O. W. Schlichter

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YELLOWSTONE-BEARTOOTH-BIG HORN REGION

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YELLOWSTONE-BEARTOOTH-BIG HORN REGION

Prepared under the direction of

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YELLOWSTONE-BEARTOOTH-BIG HORN REGION

Prepared under the direction of RICHARD M. FIELD

INTRODUCTION

By W. T. THOM, Jr.

LOCATION AND GEOGRAPHIC RELATIONS

As shown by Figure 1 the Yellowstone-Beartooth-Big Horn region is situated in south-central Montana and northwestern Wyoming. It lies at the border between the Rocky Mountains

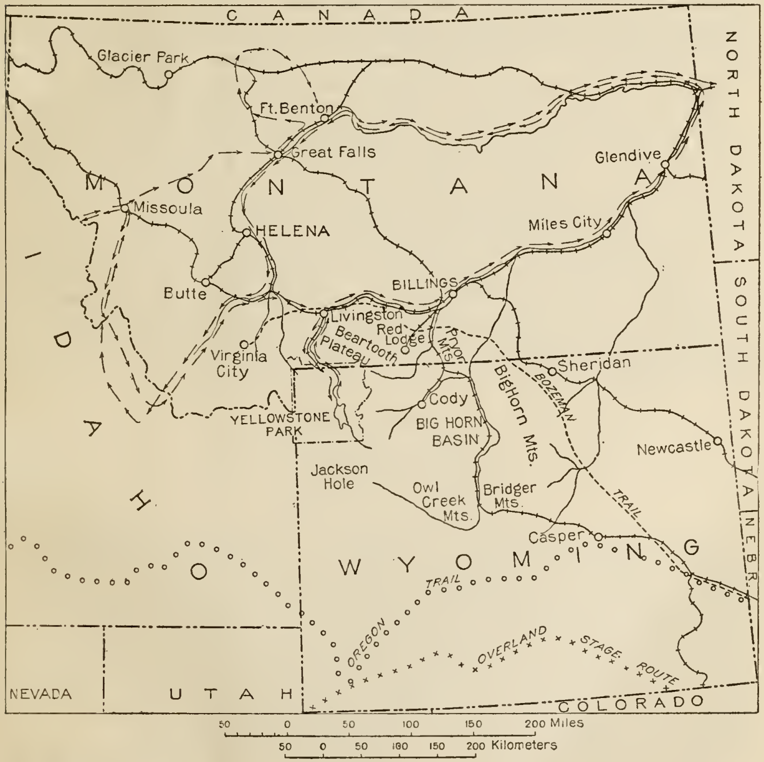


FIGURE 1.—Map showing location of Yellowstone-Beartooth-Big Horn region. Arrows show routes traversed by Lewis and Clark expedition, 1805–6

and the High Plains and is naturally divisible into the Yellowstone Park area and bordering Beartooth Plateau, the Big Horn Basin lowland, and the Pryor, Big Horn, Bridger, and Owl Creek mountain zone of uplift, which forms the eastern and southern rims of the basin.

HISTORY OF EXPLORATION

Exploration in the Yellowstone region began in 1806, when John Colter (Coulter) left the Lewis and Clark expedition near the site of the present town of Livingston and journeyed southward up the Yellowstone far enough to discover the canyon and geysers. After Colter's return Captain Clark's party followed the Yellowstone eastward, traversing the domain of the Absarokee (Crow) Indians and passing within sight of the Beartooth and Pryor Mountains and the Clark Fork Valley. In 1844 James Bridger, noted Rocky Mountain guide, described the "wonderful" springs and geysers, but his stories, like Colter's, were discredited. In 1869 Cook and Robson published an account of the wonders of the region in a Chicago newspaper.

By that time countless scouts, trappers, miners, and settlers were moving westward over the Bozeman and Oregon Trails (see fig. 1), and among these pioneers were "Yellowstone" Kelly, whose grave overlooks Billings, and "Buffalo Bill" (Col. William F.) Cody, for whom the town of Cody is named. In 1871 a party under Dr. F. V. Hayden made pioneer geologic surveys in the Yellowstone region, and as a consequence of his work the Yellowstone National Park was established by act of Congress in 1872. Since his day the work of scores of geologists and engineers has contributed to our knowledge of the Yellowstone and adjacent regions.

GEOLOGY

STRATIGRAPHY

By John G. Bartram

Sedimentary formations ranging in age from Cambrian to early Tertiary are well exposed in the Beartooth-Big Horn region, and later Tertiary eruptives, or derivatives of them, are present in Yellowstone Park and along its eastern border. (See pl. 2.) Terrace deposits of Tertiary and Quaternary age are also extensively developed in the Big Horn Basin and along the lower mountain slopes, and moraines are conspicuous in places along the Beartooth front and in the Big Horn Mountains.

The general sequence and character of the sedimentary formations of the region are shown by the accompanying table.

Sedimentary formations exposed in the Yellowstone-Beartooth-Big Horn region

[By John G. Bartram]

Age		Formation			Character	Origin
		Billings area		Cody area (with thickness)		
Tertiary.					Andesitic tuffs and flows.	Deposits that spread from the Yellowstone Park volcanic center.
	Eocene.[a]			Wind River and Wasatch, 1,000 feet (305 meters).	Red and drab clay, buff and white sandstone, and gravel beds. Strong angular unconformity near basin margins.	Flood-plain and stream deposits formed after a period of major deformation. Contains mammalian vertebrates, leaves, and invertebrate fresh-water fossils.
	Paleocene.[b]	Fort Union.		Fort Union, 3,600 feet (1,097meters).	Buff and white sandstone, with drab and green clay; some gravel beds and seams of coal. Some angular unconformity at base near basin margins.	Fresh-water sediments deposited in flood plains, swamps, etc.
Mesozoic.	Upper Cretaceous.	Lance.[a] [b]		Lance,[a] 700 feet (213 meters).	White and buff sandstone with large concretions and yellow, green, and gray clay.	Fresh-water sandstones and clays locally, with abundant dinosaur remains and fossil leaves.
		Lennep.		Meeteetse, 1,100 feet (335 meters).	Soft gray and brown shale, with gray and buff sandstone and some coal.	Marine shale rapidly grading to sandy nonmarine beds farther west.
		Bearpaw.				
		Judith River.		Mesaverde, 850 feet (259 meters).	Prominent gray and white sandstone, with gray and brown shale and coal beds near base.	Sandstone and coal series, partly marine, partly brackish water. Source in the west; thins eastward.
		Claggett.				
		Eagle.		Cody, 2,200 feet (670 meters).	Gray and black marine shale, with Eagle sandstone near top.	Marine beds deposited not far from western shore line.
		Telegraph Creek.				
		Niobrara.				
		Carlile.				
		Frontier.		Frontier, 550 feet (168 meters).	Two or more beds of gray and buff sandstone with gray and brown shale and some bentonite.	Wedges from a thicker sandstone formation to the west; thin and disappear eastward.
		Mowry.		Mowry, 200 feet (61 meters).	Hard gray siliceous shale with fish scales. Breaks in thin rectangular fragments.	Peculiar character probably due to admixed volcanic material.
		Thermopolis.		Thermopolis, 450 feet (137 meters).	Gray to black shale with one sandstone (the "Muddy") and several bentonite beds.	Marine shale and altered pyroclastic rocks derived from west.
	Lower Cretaceous.	Cloverly.		Cloverly, 128 feet (39 meters).	Upper hard sandstone, variegated shale, and lower conglomeratic sandstone, containing abundant pebbles of black chert.	Upper bed of Cloverly is basal sandstone of invading interior Cretaceous sea.
		Morrison.		Morrison, 400 feet (122 meters).	Gray to reddish or purplish shale, white sandstone, and one or more thin beds of gray limestone.	Fresh-water material that in places contains dinosaur bones and gastroliths.
	Jurassic.	Sundance.		Sundance, 575 feet (175 meters).	Blue-green shale, yellow and green sandstone, fossiliferous sandy limestone, reddish shale, and some gypsum.	Deposited in a Jurassic sea that lay mostly to the north and west of this area. Sandstones increase to south.
	Triassic.	Chugwater.		Chugwater, 975 feet (297 meters).	Bright-red sandy shale and shaly sandstone. Some gypsum beds and thin limestones.	The "Red Beds" contain few fossils. Evenly deposited material that grades into nonred marine deposits farther west. Disappears to north.
Paleozoic.	Permian.	Big Horn Mountains.	Beartooth region.	Embar, 150 feet (46 meters).	Hard, cherty limestone and one interbedded sandstone.	Permian sea was largely to west, and limestone grades rapidly into red shale and gypsum to east.
		Embar.	Phosphoria.			
	Pennsylvanian.	Tensleep.	Tensleep.	Tensleep, 180 feet (55 meters).	Quartzitic sandstone, calcareous in some layers.	Very widespread cross-bedded marine and eolian sandstone that covers large areas in the Rocky Mountains.
		Amsden.	Amsden and Quadrant (probably).	Amsden, 200 feet (61 meters).	Cherty and sandy limestone, red and purple shale, and purple sandstone.	A catch-all formation that probably contains marine beds representative of several Pennsylvanian and perhaps upper Mississippian formations.
	Mississippian.	Madison.	Madison.	Madison, 1,200 feet (366 meters).	Buff, white, or grayish limestone with some chert.	Deposited over most of the Rocky Mountain area.
	Devonian.		Three Forks.		Gray shale and limestone.	Present in western Montana and absent in eastern Wyoming and Montana.
			Jefferson.		Brownish-buff, more or less crystalline thin-bedded limestone.	
	Ordovician.	Bighorn.	Bighorn.	Bighorn, 350 feet (107 meters).	Hard, massive light-colored dolomite and limestone.	Marine limestone, difficult to distinguish from the Madison. Characteristically shows pitted weathering.
	Cambrian.	Deadwood.	Gallatin. Gros Ventre. Flathead.	Deadwood, 1,150 feet (350 meters).	Light-colored limestone, green and red shale, and a basal arkosic sandstone or quartzite.	Deposited during progressive eastward marine transgression during Middle and Upper Cambrian time.
	Pre-Cambrian.	Granite.		Granite, schist, etc.		

[a] As prepared by Bartram the Mesozoic-Tertiary boundary was placed beneath the Lance formation.—R. M. F.
[b] The United States Geological Survey classifies Fort Union formation as Eocene and Lance formation as Tertiary (?).

Sequence of events in Tertiary history of Big Horn Basin [a]

		Central part		Border belt		Absaroka-Owl Creek mountain front	
		Erosion or deposition	Deformation	Erosion or deposition	Deformation	Erosion or deposition	Deformation
Miocene.	Upper.	(?)		Basic tuff (local).		Early basic breccias (widespread).	
	Lower.	Erosion (?).		Erosion (?).		Erosion.	
Oligocene.		Erosion (?).		Erosion.		Erosion.	
Eocene.	Upper.	Erosion (?). —(?)—	—Heart Mountain overthrust—	Erosion. —Unconformity ?—	Normal faults (local). —Heart Mountain overthrust—	Erosion. —Unconformity ?—	Normal faults (local). —Heart Mountain overthrust.
	Middle.	Andesitic tuffs (local).		Andesitic tuffs (local). —Conformity—		"Early acid breccia" (local). —Conformity—	
	Lower.	Tatman. Lost Cabin. } Wind River. Lysite. Gray Bull (Wasatch) beds. — Subdivision of Sinclair and Granger. —Local unconformity—	Local depression.	Border Wasatch (local). Erosion. —Unconformity—		Border Wasatch (local). Erosion. —Unconformity—	
Paleocene.		Clark Fork beds. Tiffany-Bear Creek. } Fort Union. Torrejon. Puerco	Depression.		Local normal faults. Small folds. Local uplift.		Extensive normal faults. Large folds. Widespread uplift.
			Widespread sinking.	Fort Union. —Local unconformity—	Widespread sinking.	Fort Union (?). —Local unconformity—	Sinking (?).
			(?)		Local warping.		Local warping.
Cretaceous.		Lance.	Sinking.	Lance.	Sinking.	Lance (?).	Sinking (?).

[a] Hewett, D. F., The Heart Mountain overthrust, Wyoming: Jour. Geology, vol. 28, p. 456, 1920.

REGIONAL STRUCTURAL RELATIONS

By W. T. Thom, Jr., R. T. Chamberlin, and W. H. Bucher

Great geologic interest attaches to the Yellowstone-Beartooth-Big Horn region because of its critical and significant position in terms of continental structure. Here, as shown by Plate 3, two major deformational zones converge as they are traced northward from the southern United States and simultaneously undergo a notable transformation along a line which may be drawn from Butte, Montana, to the north end of the Black Hills. Along this line the Black Hills, Big Horn, and Pryor uplifts terminate northward and the margins of the western (thrust) zone show notable changes of direction.

Of the two major deformational zones, the western, which borders the Colorado Plateau region, is characterized particularly by strong overthrust faults, which are observed in many places from southernmost Nevada northward to eastern Idaho, western Wyoming, and southwestern Montana. These faults characteristically dip westward at low angles, and many of them involve a movement from west to east of at least 20 miles (32 kilometers).

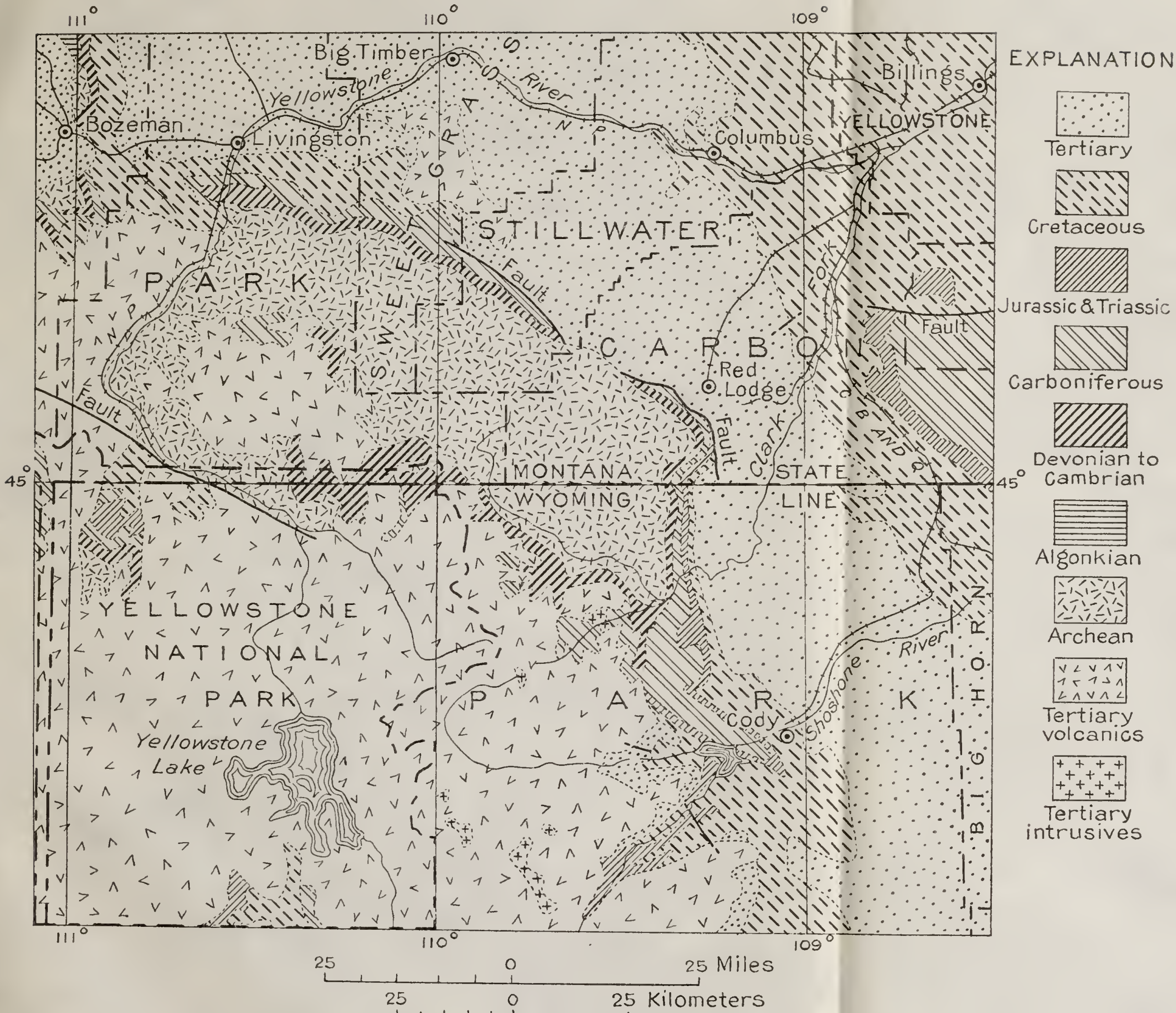
In contrast with this western belt of strong overthrust faulting is an eastern zone of deformation featured by a number of elongate mountain uplifts, strung together to give a peculiar pattern of arcuate lines, which curve northwestward to meet the western zone of thrusts at several points and finally disappear in central Montana. Curving outward from the main chain of the Rocky Mountains on their east side are two prominent arcs—(1) the Hartville uplift, extending northeastward and running into the Black Hills, which swing around into a northwesterly trend; (2) farther northwest, the Owl Creek-Bridger, Big Horn, and Pryor Mountains, which form a more regular sicklelike bend. These two belts form roughly concentric arcs, concave toward the northwest, which partially inclose the broad Powder River and Big Horn Basins. (See pl. 3.)

The eastern Cordilleran belt is characterized by localized uplifts, mostly with crystalline cores, developed as crustal folds. The dynamic peculiarity is the localization of the regional stress in such a way as to produce separate massifs, each behaving in individual manner. Thus, for example, the main massif of the Big Horn Mountains is strongly asymmetric, leaning toward the Great Plains with steeply upturned strata on the eastern flank and evidently with local overthrusts toward the east. Near the north end, however, the symmetry is reversed in places, with gentler dips toward the east and with a minor crystalline massif on the west side, lined with steeply dipping foothills and even with local overthrusts toward the west.

North of central Montana the Rocky Mountain front becomes structurally simpler and more regular, comprising for 1,000 miles (1,600 kilometers) into Canada a single orogenic belt, folded in the central and western portions, thrust faulted in the eastern portion, and separated from the Great Plains by a line of strong overthrusts and subsidiary imbrication. East of the Rocky Mountains proper in central and northern Montana the Great Plains are broken up by isolated mountain domes, mostly related to igneous intrusions. These barely reach into Canada.

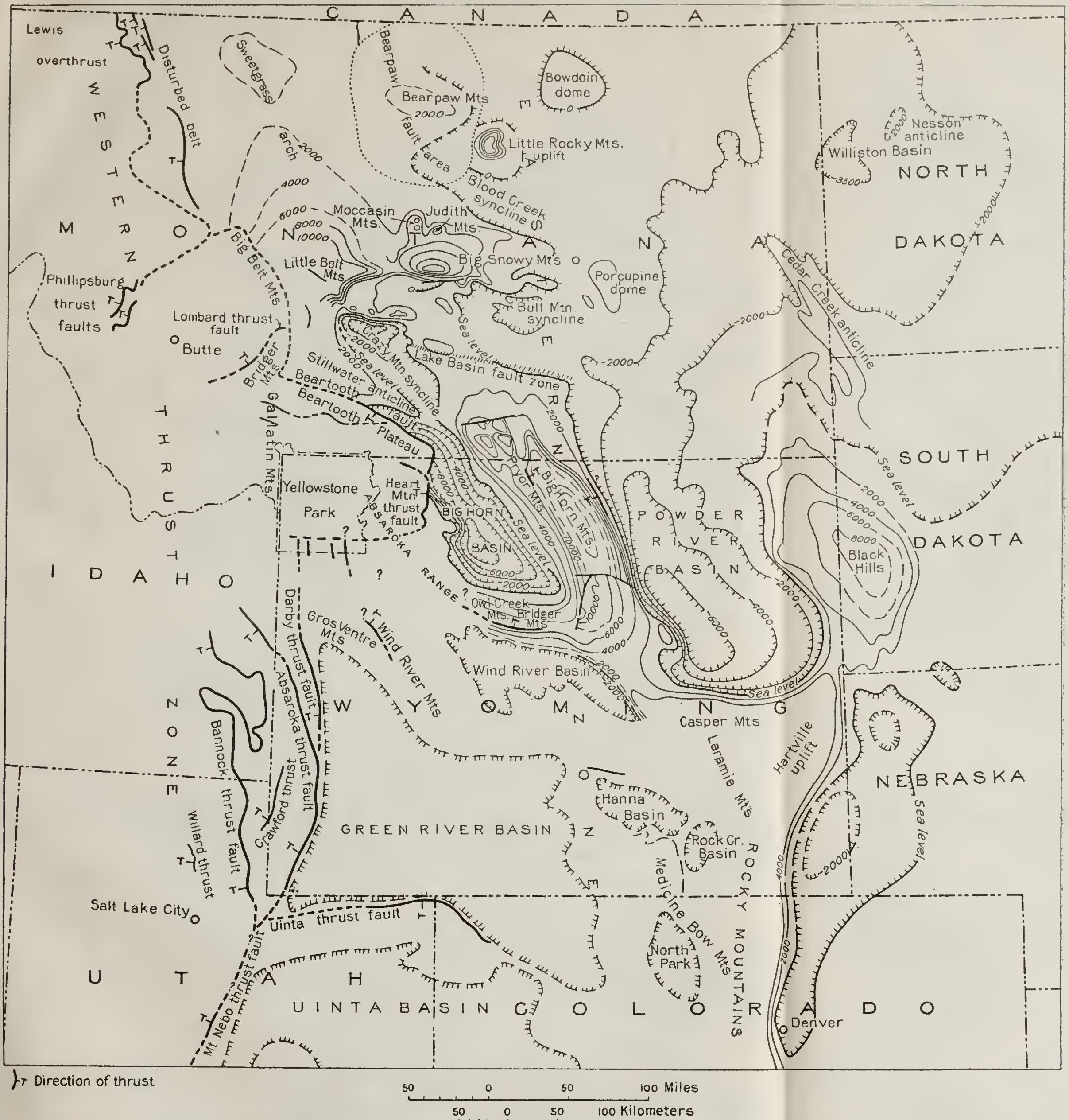
The ranges and "massifs" of the eastern zone bear witness to the simultaneous operation of both tangential thrust and vertical uplift, which together have caused thrust faulting; zones of en échelon faults; wedge elevation and "pinching"; block rotation; and, locally, laccolithic intrusion. Evidence is also accumulating to show that both the local and the regional features of this zone owe their character partly to a system of pre-Cambrian lines of weakness in the basement crystalline rocks which trend slightly south of east, and partly to the Cretaceous-Eocene (Laramide) folding and fracturing, trending about N. 15° W., which has been superimposed upon the older tectonic pattern.

The major orographic feature of the Yellowstone-Beartooth region, then, is a huge thrust slice, segmented by transverse faults and modified by subsequent erosion. This feature is best exposed along the east front of the Beartooth Plateau and in the vicinity of Heart Mountain. The local mountain uplifts and subsidiary features of the areas lying east of this thrust slice are due to wedge uplift and compression (Big Horn Mountains), to block elevation and rotation (Pryor Mountains, Big Coulee-Hailstone dome, Cedar Creek anticline, etc.), or to progressive accentuation of the Big Horn Basin depression, with consequent production of a system of anticlinal folds paralleling the basin rims. A noteworthy belt of en échelon faults also occurs in the Lake Basin region, just north of Billings, and a similar though less striking and less extensive zone is developed along a part of the Stillwater-Dry Creek anticlinal axis, indicating a relative eastward shift of the south walls of basement faults underlying the belts of en échelon faulting in the surface sedimentary formations. The smaller and more typically anticlinal folds—for example, that of Elk Basin (see fig. 8)—show notable epi-anticlinal fault systems, due presumably to tension incident to the vertical arching of a double plunging fold. Most of the anticlines are, however, strongly asymmetric, and much evidence is available suggesting that they owe their form to the draping of the sedimentary beds over the edges of rotated or tilted blocks of the basement crystalline rocks.



GEOLOGIC MAP OF PARTS OF MONTANA AND WYOMING

From Geologic map of the United States, compiled by G. W. Stose.



SKETCH MAP SHOWING STRUCTURAL SETTING OF YELLOWSTONE-BEARTOOTH-BIG HORN REGION

Contours show approximate altitude of Dakota or Cat Creek sandstone.

GEOLOGIC HISTORY

By Arthur Bevan, Eliot Blackwelder, N. M. Fenneman, R. M. Field, and W. T. Thom, Jr.

During later pre-Cambrian time most if not all of eastern Montana and Wyoming seems to have been emergent. The pre-Cambrian granites were exposed by erosion and reduced to an almost plane surface over very wide areas, and the débris was deposited in the geosynclinal trough in Idaho and western Montana, in which many thousand feet of clastic sediments and limestones (Belt series) accumulated during pre-Cambrian time. Progressive filling of this trough led to the gradual transgression of the Cambrian seas across the beveled granite surface to the east, and the Cambrian normally consists of a basal conglomerate of local derivation and overlying calcareous shales and limestones. Repeated minor oscillations of sea level are indicated by "flat-pebble" intraformational conglomerates in the Upper Cambrian, and other oscillations of greater magnitude and duration are indicated by a number of hiatuses between the essentially parallel formations that make up the local Paleozoic section. Some differential vertical movements of large crustal segments are indicated by differences between the stratigraphic sections of southern and central Montana, but no appreciable tilting or folding is known to have occurred in southern Montana and northern Wyoming until about the middle or end of the Mississippian period.

After the repeated flooding of the region by widespread and shallow seas in pre-Pennsylvanian time a restriction of the sea in the Pennsylvanian epoch was accompanied by the development of the very extensive and strongly cross-bedded Tensleep sandstone, which is perhaps largely of dune character. In Permian time the local area, like the "Red Beds" region of western Texas and southeastern New Mexico, was one in which red shale and gypsum were deposited while limestone was being laid down in the adjacent sea to the west.

Red beds were also developed during the Triassic, while marine sediments of normal character were being laid down a little farther west. The Chugwater formation, from its gypsum and chemically precipitated limestone content and its regular bedding, appears to be prevailingly a saline-lagoon accumulation.

The sea withdrew in later Triassic time but again covered the region in the Upper Jurassic. The first crustal stirrings premonitory of the coming Laramide revolution are indicated by the red material locally reworked from the Chugwater into the basal Jurassic, which is normally gray or green. Also notable variations in thickness and composition of the Sundance and

Morrison formations show that the Big Horn mountain arch was incipient in pre-Cretaceous time, in sympathetic response to the far stronger movements which were taking place at that time in regions farther west.

The eastward movement of the wave of deformation and intrusion affecting the Cordillera in the Mesozoic and early Tertiary is shown by variations in the thickness and character of the Cretaceous deposits. Recurrent uplifts of positive elements, especially in regions to the west, afforded fresh increments of clastic sediment which were deposited in the progressively depressed seaway and embayments of the eastern Cordillera. During late Upper Cretaceous time the shore line lay near meridian 110° and local variations in both thickness and character of the late Upper Cretaceous and Cretaceous-Eocene transition beds indicate the recurrence of progressively stronger compressive movements, which in turn gave rise to moderate uplift of the present mountain chains in post-Lance time and notable uplift after Fort Union time. It has been considered that the Laramide deformation in the Yellowstone-Beartooth region culminated in profound thrust faulting, which occurred after the middle Eocene and before the Oligocene. As a result of the Laramide orogeny there was a transition from the marine to the continental-basin type of deposition, which has ever since been characteristic of the Rocky Mountain region.

Downcutting of the region then ensued, and apparently by Oligocene time the whole of eastern Montana and Wyoming not already planed off by erosion during the Cretaceous and early Eocene was reduced nearly to base-level. Miocene uplifts, accompanied by prodigious volcanic activity in the Yellowstone Park region, initiated a new period of erosional down cutting.

Recurrence of upwarping of the mountain areas in late Tertiary and Quaternary time is indicated by the presence of at least one upraised peneplain, the subsummit upland; by the presence of stripped surfaces where resistant beds, such as the Tensleep sandstone in the Pryor Mountains, are overlain by weaker ones; and by extensive systems of terraces developed along the Beartooth and Big Horn Mountains on the weaker rocks of the Great Plains and Big Horn Basin. The highest terrace level, embracing Polecat Bench and some of the higher mesas, was a widespread peneplain developed 3,000 feet (914 meters) or more below the older peneplain preserved on top of the Beartooth Range. The basin of each stream issuing from the mountains has its own independent system of lower terraces. Valleys in the mountains and to some extent the uplands have also been materially modified by repeated glaciation.

YELLOWSTONE NATIONAL PARK

By Richard M. Field

INTRODUCTION

The Yellowstone National Park occupies the northwest corner of Wyoming, slightly overlapping Montana to the north and Idaho to the west. (See fig. 1.) The park is 62 miles (100 kilometers) long and 54 miles (87 kilometers) wide (area 3,348 square miles, or 8,672 square kilometers) and is under the control and supervision of the National Park Service of the Department of the Interior. The Yellowstone is probably the most celebrated of the national parks, owing to its hot springs and geysers. The rocks of the park are largely volcanic, including one of the most extensive known exposures of acid flows, as well as a great series of volcanic agglomerates, of Tertiary age. The Tertiary and Quaternary stratigraphy and physiography are not only unique in North American geology but also bear an important relation to the contiguous areas.

TOPOGRAPHY [1]

The central portion of the Yellowstone Park may be described as a broad volcanic plateau between 7,200 and 8,300 feet (2,195 and 2,530 meters) above sea level, with an average altitude of nearly 8,000 feet (2,438 meters). (See pl. 1.) Surrounding this plateau on the south, east, north, and northwest rise mountain ranges with culminating peaks and ridges standing from 2,000 to 4,000 feet (610 to 1,219 meters) above the general level of the inclosed area. The Gallatin Range, which shuts in the park on the north and northwest, is a bold range extending from Mount Holmes, at the south end, far northward into Montana, where it presents a sharply defined ridge along the west side of the Yellowstone Valley. Electric Peak, on the northern boundary of the park, the highest point of the range, attains an altitude of 11,100 feet (3,383 meters). South of Mount Holmes the lavas that form the Madison Plateau stretch southward beyond the limits of the area shown on the map. Teton Range, on the south, forms one of the most prominent geographic features of the northern Rocky Mountains. The eastern wall of this mountain mass rises nearly 7,000 feet (2,134 meters) above Jackson Lake, which lies immediately at its base in the open valley of the Snake River. Northward the ridges of the Tetons fall away abruptly and terminate just south of

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[1] Largely from U. S. Geol. Survey Geol. Atlas, Yellowstone National Park folio (No. 30), 1896.

the Yellowstone Park boundary. To the east of the Tetons, across the broad valley of the Snake, are the Gros Ventre and Wind River Ranges.

Forming an unbroken barrier for 80 miles (129 kilometers) along the entire east side of the park stretches the Absaroka Range, so called from the Indian name of the Crow Nation (Absarokee). At its south end the range is topographically closely connected with the Wind River Range by the Wind River Plateau.

Across the elevated plateau inclosed by these mountains lies the Continental Divide, which crosses the park from southeast to northwest. Following the top of Two Ocean Plateau, it skirts the northern escarpment of Flat Mountain, winds among the undulating low ridges lying between Yellowstone and Shoshone Lakes, and thence, sweeping around the streams running into Shoshone Lake, crosses the Madison Plateau and leaves the park a short distance southwest of the Upper Geyser Basin.

Four large rivers originate in this region. The Snake River drains the south and west side of the great divide, and the Yellowstone, Madison, and Gallatin flow to the east and north. A large part of the central plateau is drained by the Firehole and Gibbon Rivers, which unite to make the Madison. The Madison River, running nearly due west, cuts a deep gorge through the Madison Plateau and, after a winding course, pours its waters into the Missouri at Three Forks. Within the park the Gallatin drains a much smaller region than any other of the large rivers, being restricted to the western slope of the Gallatin Range, with a drainage area of less than 80 square miles (207 square kilometers). Like the Madison, it empties into the Missouri at Three Forks. Across the Park Plateau and the Absaroka Range the country presents an unbroken mountain mass over 75 miles (121 kilometers) in width, with an average altitude unsurpassed by any other area of equal extent in the northern Rocky Mountains.

GEOLOGY AND GEOMORPHOLOGY

SUMMARY OF GEOLOGIC HISTORY

A summary of the geologic history of the Yellowstone Park is given in the following table:

Geologic history of Yellowstone Park

Quaternary.	Recent.	Removal of major portion of valley basalts and lake sediments. Reexcavation of the Grand Canyon of the Yellowstone and Lamar Valley. Local valley glaciation.		
		—Erosion—		
	Pleistocene.	Glaciation, piedmont type. Till and glacial-lake deposits (Bull Lake epoch?).		
		Basalt "valley" flows and associated lake sediments. Canyon filled to brim; rim conglomerate.		
		—Erosion—		
		"Trachytic rhyolite" flows and associated lake sediments and canyon conglomerates.		
		—Erosion—		
		Basalt "valley" flows and associated lake sediments and canyon conglomerates.		
		Intrenchment of Yellowstone River. The Grand Canyon cut to approximately present depth of the Yellowstone.		
		Uplift and rejuvenation.		
Tertiary.	Pliocene.	Development of local base-level. Removal of 600 ± feet (183 meters) of rhyolite.		
	Miocene.	Extrusion of rhyolite and associated acid flows and volcanic rocks.		
	Oligocene.	Extensive erosion. Development of mature surface. Mount Washburn a monadnock.		
	Eocene.	"Basic" breccias and agglomerates. Pinyon conglomerate.	Andesitic breccias with basic intrusives and flows.	
		"Acid" volcanic breccias.		
Mesozoic.	Upper Cretaceous.	Montana group undivided. Thick sandstone above and shale below.		
		Colorado group undivided; 2,000 feet (610 meters) of shale with some limestone.		
	Lower Cretaceous.	Kootenai formation; 250 ± feet (76 meters). Shale, quartzite, and sandstone. Mapped under color for Cloverly formation on Wyoming State map.		
	Jurassic.	Ellis formation; 500 ± feet (152 meters). Limestone, marl, and shale.		
		(?)		
	Triassic.	Teton formation; 200–400 feet (61–122 meters). Shale, sandstone, and cherty limestone.	(?)	
			Limestone.	
			Red sandy shale.	800 feet (244 meters).
			Shaly brown limestone.	
Paleozoic.	Permian.		Phosphoria formation; 100–340 feet (30–104 meters). Shale, quartzite, chert, and phosphate rock.	
	Pennsylvanian.	(?)		
		Quadrant quartzite; 200–425 feet (61–130 meters). Quartzite, sandstone, some limestone, and shale.		
	Mississippian.	(?)		
		Madison limestone; 1,300–1,600 feet (396–488 meters). White or cream-colored limestone above; dark-colored limestone below.		
	Upper Devonian.	Threeforks limestone; 170–250 feet (52–76 meters). Gray cherty limestone.	Darby formation in Teton Range.	
	Middle Devonian.	Jefferson limestone; 110–300 feet (34–91 meters). Dark crystalline limestone.		
	Upper Ordovician.	Big Horn dolomite; 200–300 feet (61–91 meters). Massive white to buff dolomite.		
	Upper and Middle Cambrian.	Gallatin limestone; 110–400 feet (34–122 meters).	Massive limestone, shale, and "edgewise" conglomerates.	
			Calcareous and argillaceous shale.	
			Dark mottled limestone; 100–500 feet (30–46 meters).	
	Middle Cambrian; 700–750 feet (213–229 meters).	Gros Ventre formation. Shale and thin limestone near top.		
		Flathead quartzite. Conglomerate at base.		
Algonkian.		Sheridan quartzite.		
Archean.		Granite and gneiss.		

104787—32. (Face p. 8.)

A fairly complete geologic history is illustrated within the confines of the park. The pre-Cambrian and Paleozoic formations are well exposed north of Junction Butte. (See fig. 2.) A part of the Mesozoic section is exposed on the west side of Mount Everts, a view of which can be obtained from the Mammoth Hot Springs terrace. The Tertiary history of the park is

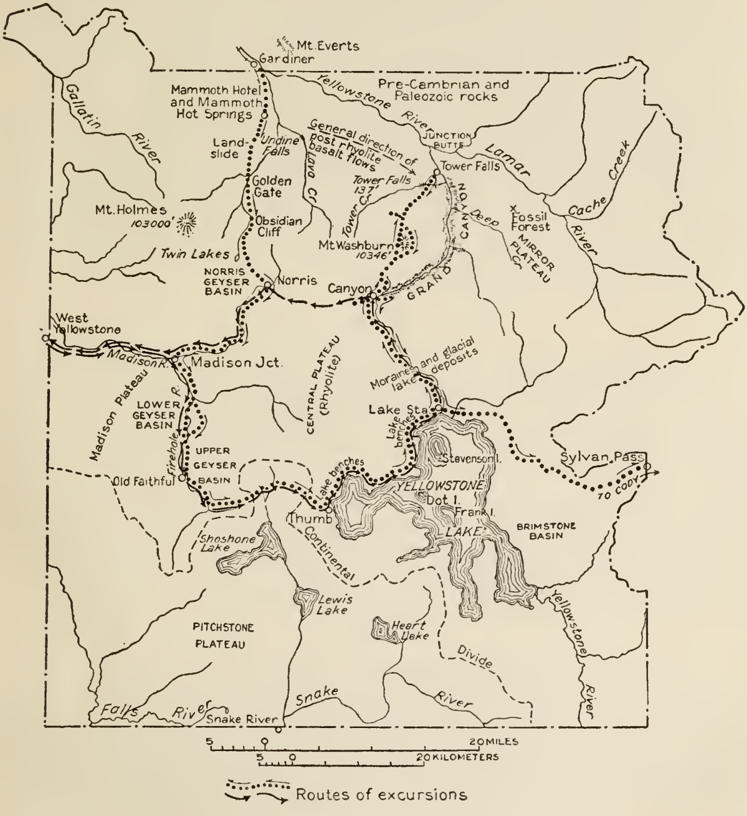

FIGURE 2.—Sketch map of Yellowstone Park showing routes of the excursions. Dots indicate route of excursion C–2; arrows, C–1

largely volcanic and erosional, and the origin of the geomorphic features is closely related to the volcanic history. According to Hague (3)[2] the later basalts, such as those exposed at Junction Butte and Tower Falls, are interbedded with the rhyolites. Recent investigations have shown, however, that the later basalts

[2] Numbers refer to bibliography, p. 23.

exposed in or near the valleys of the Yellowstone and Lamar Rivers are postrhyolite. This interpretation has seriously modified the geologic and geomorphic history of the park.

VOLCANIC PERIOD

Throughout post-Cretaceous time the Yellowstone Park region has been marked by great volcanic activity. Enormous volumes of erupted material—flows, breccias, and agglomerates—were ejected during the Tertiary period. Active volcanoes surrounded the park on the east, north, and west. The Absaroka Range is carved from an enormous thickness of ejectamenta and interbedded sediments, and the depressed basin lying between the encircling ranges was filled some 600 feet (183 meters) above its present altitude by flows of rhyolite and associated eruptives which form the basement rocks of much of the present plateau. After the Cretaceous-Eocene revolution the region was subjected to denudation on a profound scale.

EOCENE

During the Eocene epoch the region was covered by a great thickness of agglomerates, breccias, and mud flows, chiefly andesitic breccias. The total thickness of these deposits has not been definitely determined, but they are more or less continuously exposed for some 30 miles (48 kilometers) east of Sylvan Pass and dip at a relatively low angle. Fossil plants from sediments intercalated in the older breccias show these beds to be of Eocene age (according to F. H. Knowlton), as is the flora of the somewhat younger "basic" breccia.

OLIGOCENE

This period of intense volcanic activity was followed by a long period of erosion, with little or no volcanic activity, which probably lasted throughout Oligocene time. The original somewhat irregular surface was reduced to a relatively mature plain surmounted by monadnocks, such as Mount Washburn, which probably was never covered by the rhyolite flows.

MIOCENE

Upon this relatively old erosion surface, in Miocene time, a great series of acid lava flows was poured out. It is probable that the rhyolite was erupted at intervals along several fissures. Daly has suggested that the rhyolite may be explained by the foundering of the roof of a batholith. The field evidence suggests widespread and progressive fissure eruptions.

PLIOCENE

The extrusion of the rhyolite was followed by another long period of erosion, which continued through a good part if not all of Pliocene time. Nowhere in the region does this rhyolite show the original chilled surface except possibly at Obsidian Cliff. (See p. 24.) Furthermore, the rhyolite is found between the spurs of Mount Washburn several hundred feet above the general level of the plateau. In places near the eastern border of the park the rhyolite rises to an altitude of about 10,000 feet (3,048 meters), from which it descends to the Lamar Valley, forming a capping to the long spurs that project into that valley on its east side. The difference in altitude between different portions of the rhyolitic surface, the difficulty of accounting for the slope of the rhyolite surface between the spurs of Mount Washburn (except on the supposition that the surface had once

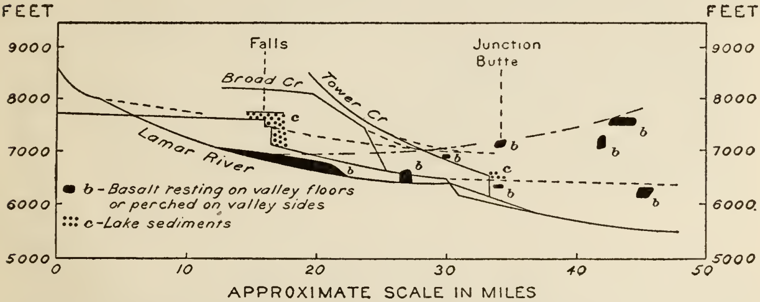

FIGURE 3.—Profiles of the Yellowstone River and tributaries in relation to lava-flow relics. Solid lines, present profiles; even dashed lines, profiles prior to rejuvenation; uneven dashed line, maximum thickness of the valley flows. (After O. T. Jones).

stood at a considerably higher level), and the evidence from some of the tributaries of the canyon that they had attained a late stage of development in a cycle of erosion preceding the cycle which initiated the canyon (see fig. 3) suggest that the present rhyolite surface had been attained as the result of prolonged subaerial denudation before the canyon itself was cut.

PLEISTOCENE

The cycle of erosion during which the Yellowstone Canyon was cut began some time after the eruption of the rhyolite, approximately at the boundary between Pliocene and Pleistocene time. Between the upper and lower falls there is evidence that when the river began to cut down it was in the meandering state (9, p. 275).

The canyon cycle was ended by a series of basaltic and trachytic lava flows separated by erosion intervals (see table, p. 8) which progressively dammed the mouth of the canyon, until after the last (basaltic) flow the canyon was completely dammed and filled to the brim with lacustrine sediments. Good exposures of these sediments occur in abandoned meanders at several places in the canyon walls. A section at Red Rock, near Canyon Camp, includes in descending order the following strata:

	Feet	Meters
15. Blue muddy yellow-weathering laminated silt, in part flaggy	65	20
14. Coarse cross-bedded conglomerate	10	3
13. Very coarse, strongly cross-bedded pebbly sandstone, with silty bands	12	3.7
12. Coarse cross-bedded yellow sand with small brown spots	10	3
11. Coarse sand with pebbles, interbedded with silty layers	7	2.1
10. Fine-grained well-laminated blue muddy silt	20	6
9. Conglomerate, coarse at base, finer upward	2½	.8
8. Fine-grained cross-bedded blue muddy sand with green nodules	20	6
7. Coarse cross-bedded blue sand	3	.9
6. Medium-grained conglomerate	1½	.5
5. Muddy silt, slightly cross-bedded	6	1.8
4. Cross-bedded sandstone with pebbles	2½	.8
3. Breccia	4½	1.4
2. Coarse cross-bedded sandstone with pebbles	3½	1.1
1. Coarse conglomerate	20	6
	187½	57.1

The base of the section is estimated to be about 350 feet (107 meters) below the rim of the canyon, which is here about 700 feet (213 meters) deep. Combining the exposures in the neighborhood of the Yellowstone Falls and at Red Rock gives an almost complete record of the lacustrine sediments that once occupied the canyon, from the rim to a level within 50 feet (15 meters) of the river bed.

Soon after the eruption of the valley flows and the consequent damming and filling of the canyon the region was glaciated, the ice moving nearly due south from the direction of the Snowy Range, northeast of Yellowstone Park, and across the buried Lamar Valley and canyon toward the Park Plateau. Erratics, till, and glacial-lake deposits lie on both the rhyolite and the latest beds of the lacustrine epoch (rim conglomerate).

Since the last glacial epoch exceedingly active stream erosion has removed nearly all traces of the preglacial valley and lacustrine sediments and has also removed the greater part of the valley flows, remnants of which remain perched on the valley sides or cover parts of the valley floors. To those who may be skeptical as to the amount of lava which has been removed from the valleys of the Lamar and Yellowstone Rivers since Pleisto-

cene time it may be stated that several other examples are known in Idaho, Arizona, and elsewhere in the West and Southwest.

There is good evidence of late valley glaciation, especially in the valley of the Lamar River.

HOT SPRINGS OF YELLOWSTONE PARK

By E. T. Allen

The hot springs of the Yellowstone Park are scattered across the Central Plateau, within an area 60 miles (97 kilometers) from north to south and about 30 miles (48 kilometers) in maximum breadth. There are about 100 clusters, including about 3,000 springs. These range in size from insignificant spring holes to hot pools 350 feet (107 meters) or more in diameter, with a maximum depth accessible to the plummet of about 70 feet (21 meters). In large numbers of these springs the surface temperatures are close to the boiling point, in many actually surpassing it. They occur at altitudes ranging from 6,200 to 8,500 feet (1,890 to 2,590 meters), but the principal groups are found in localities where the boiling point of pure water varies with atmospheric pressure around 93° C. Inasmuch as most of the waters contain less than 1.5 grams of dissolved mineral matter to the liter, much of which is usually silica, the boiling point is seldom raised more than 0.02° to 0.03° C. from this cause. Gases, on the other hand, may lower the boiling point considerably.

Relation of springs to rocks.—The springs are all associated with rhyolite flows that cover the central portion of the park; most of them issue from this rock, but a few are more directly related to limestone (Mammoth Springs) or to breccias made up of basaltic or andesitic fragments (Calcite and Washburn Springs).

Geologic age.—Glacial drift on the summit of Terrace Mountain, a travertine deposit of hot springs now extinct, proves that thermal activity in that locality antedated at least the latest ice invasion. Equally clear proofs are not found elsewhere, but the uniform character of the rhyolite body, as shown by the lack of weathering between the flows, and the absence of distinct stages recorded in the topographic relief, as well as the fact that rock sculpturing by hot water on the rhyolite plateau is in many places obviously ancient, convinced Arnold Hague that the thermal activity was all of the same age, following closely the cessation of the Pliocene rhyolite flows and lasting ever since.

Genesis of the hot springs.—Measurements by the Geophysical Laboratory in the summer of 1930 proved that the aggregate

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discharge of hot water in the park amounted to about 110 cubic feet (3,115 liters) a second, an amount less than the cold outflow of a single drainage basin in the park. The rainfall, though only about 20 inches (0.5 meter) annually, is thus abundantly able to supply the hot springs many times over. That it does supply them is indicated by the fact that the hot discharge is in harmony with the topography. Spring groups that occur in drainage basins where the catchment area is large have an abundant flow; those found on steep slopes, in narrow ravines, or in shallow depressions have a very meager flow.

On the other hand, a body of fact almost equally cogent clearly links the springs to volcanism as one of its distinctive phases. Gases seen bubbling through the water of many springs are volcanic gases. They escape from the fumaroles or steam holes prevalent in the volcanic regions, and the expansive force of such gases is believed to be the cause of the rise of lavas in craters and of the disastrous explosions associated with certain active volcanoes.

The original gas consists mostly of steam, by which practically all the heat of hot springs is transported and by the condensation of which most of the heat is given up.

In many active areas hot springs and steam vents occur together. Which of the two shall form seems to be dependent upon the relative amounts of ground water and of volcanic steam supplied to a particular spot and upon the temperature of each. When the steam starts out from the magma it is doubtless superheated, and sometimes it remains so when it reaches the surface. Eleven superheated vents, the hottest of which has reached 138° C., have been found in the Yellowstone Park. It is therefore believed that the portion of the water consisting of condensed steam must be of magmatic origin. It probably nowhere amounts to more than 15 per cent of the total volume.

Though steam is the most abundant of the volcanic gases, there are others that play a part in hot springs economy. Carbon dioxide and the oxidation product of hydrogen sulphide (sulphuric acid) are unquestionably active agents in the alteration or disintegration of rock, and both of these gases, and perhaps also hydrochloric acid and hydrofluoric acid, are instrumental in bringing mineral matter into solution in the hot spring waters.

Types of hot areas.—Three principal types of hot springs and three corresponding types of hot areas are found in the Yellowstone Park. The distinctive differences in the two fundamental types are caused by differences in the volume of ground water and differences in the supply of sulphur brought up by hydrogen

sulphide with other gases. The third type of hot ground is conditioned by the accidental association of limestone with the hot waters. The three types of hot areas follow.

1. The typical geyser basins occupy valley floors or well-marked depressions at the foot of slopes that supply an abundant volume of ground water. The clear alkaline waters are characterized by dissolved silica and sodium salts, though potassium, lithium, and calcium salts are also found in small amounts. The compounds of the metals that occur most abundantly in the alkaline waters are bicarbonates (with or without carbonates) and chlorides. Fluorides, borates, sulphates, and traces of arsenates are also characteristic.

Practically the only deposit in this type of hot ground is siliceous sinter, which not only lines the springs and their outlet channels if they overflow but builds up borders about them several inches in height, which here and there grow into mounds, cones, or irregular structures. The sinter covers also more or less of the space between the springs, in places extending over an area of many acres. Where the water has been diverted to other courses the surface layer of the sinter disintegrates into sand and gravel.

No free sulphur is found in these areas, and only slight amounts in the form of sulphates, etc., occur in the waters.

Thermal activity as measured by the volume of water near the boiling point is high. About 73 per cent of the total discharge of hot water in the Yellowstone Park is of this type. Geysers and superheated waters are characteristic of these alkaline areas, but on account of the large volume of ground water superheated steam vents are absent.

Geyser studies by the Carnegie Institution have been pursued here for many years—on the circulation of the water, the characteristics of intermittent discharge, the altitude of eruptions, the volume of the gases escaping, the variability of action, and past history. From these studies it is concluded that geysers are a phenomenon considerably less permanent than hot springs, and that most geysers, especially large ones, are subject to fluctuations in their action which are certainly not seasonal, though not yet satisfactorily accounted for. One cause of permanent change in geysers that now seems probable, though hitherto unsuspected, is the variation in steam supply and therefore in heat supply, due to the gradual opening up of seams by hot water. In four different localities (not in the Upper Basin, however), where geysers have been seated in clay or loose material, they have proved short-lived.

Bunsen undoubtedly grasped the secret of geyser action in the boiling of water at some depth below the surface, but even

as modified to-day the generally accepted Bunsen theory does not account for the intermittent overflow, accompanied or unaccompanied by eruptions, nor for the long periods and high discharge of certain great geysers. Neither is it in good accord with the circulation of water so generally observed in geyser wells. Only a subterranean chamber connected with the wells is thought adequate to account for these facts.

2. Sulphate areas occur on mountain or hill slopes, in ravines, or in shallow depressions. The springs in these areas are generally small and shallow and are fed, apparently, by a circulation of comparatively little depth. The waters contain sulphates of the common rock metals, with only traces of chlorides and fluorides. As a rule they contain some free sulphuric acid. These barren tracts are covered with residual quartz sand derived from decomposed rhyolite and are dotted with springs, commonly more or less choked with mud, consisting of opal and kaolinite and in some places precipitated sulphur. Abundant gases are usually bubbling through the water. The temperatures in such springs are usually lower than temperatures in alkaline areas. Superheated waters are never found in this type of ground, and geysers rarely, but it is in such places, low in ground waters, that superheated steam vents occur. Numerous as these areas are, their aggregate flow amounts to only a trifle more than 1 per cent of the entire hot discharge of the Yellowstone Park.

3. Only two or three travertine areas showing thermal activity now exist in the park. The combined discharge of waters of this class is about 22 per cent of the aggregate hot outflow. The essential conditions for the development of such areas appear to be a continuous supply of volcanic gases rising through water and a body of limestone with which the water comes into contact. Travertine deposits distinctive of these localities usually crystallize in the form of calcite, though aragonite also occurs.

Several areas constituting a subdivision of this group are distinguished by both travertine and silica deposits, occurring at some points in mixtures but usually separate. The clear alkaline waters are much lower in calcium and much higher in silica than those of the type. Areas of this class possess abundant supplies of gas, but none of them are associated with limestone outcrops. From the low concentration of calcium in the waters it is surmised that this element, like the rest of the mineral matter, is derived from the rhyolite through which the waters percolate. On the other hand, the fact that the waters are higher in calcium than those of type 1 may be explained by the greater abundance of gas, which prevents the precipitation of calcite below the surface. The higher calcium concentration may also

be due partly to a locally higher percentage of lime in the rhyolite. Analyses show that the rhyolites of the Yellowstone vary greatly in lime.

Mixed waters.—Although neither the hot waters nor the hot areas are usually of perfectly pure type, as a rule one kind of spring predominates. In a few localities a considerable percentage of the mineral matter in the water is derived from both the deep and the shallow circulation. Such mixed waters aggregate about 4 per cent of the total hot discharge in the park.

HOT AREAS EASILY ACCESSIBLE TO THE TOURIST

1. The *Mammoth Hot Springs* lie on the border of a drainage basin, plentifully strewn with glacial débris, none of which covers any of the present spring deposits. Drift occurring above old travertine deposits high upon Terrace Mountain, to the west, proves that those deposits are preglacial. The nearest outcrop of rhyolite underlies this travertine. The temperatures here are much lower than they are in most other hot springs of the park—never quite 75° C.; perhaps the volume of magmatic steam per unit volume of ground water is low. Only a small part of the ground water discharged in the basin comes to the surface in the Mammoth Springs; 10 to 20 times as much is discharged by the Hot River, 1½ miles (2.4 kilometers) to the northeast. Careful weir measurements during recent years have shown that the aggregate discharge of hot waters in this area ranges from about 1.1 to 3.0 cubic feet (0.03 to 0.08 cubic meter) a second. A prolonged drought has prevailed in the park since the measurements were begun, but it does not yet appear how far the variations are due to that cause and how far to the disappearance of springs within the area. The waters of the Mammoth Springs are high in calcium, bicarbonate, and sulphate, moderate in chloride, and low in silica. The water of the Hot River is similar except that its concentration is only about 6 per cent of that of the Mammoth Springs, and its temperature only 46° to 58° C.

A phenomenon of importance in connection with the deposition of travertine at Mammoth is the large amount of volcanic gases escaping from the springs. All magmatic gases contain far more carbon dioxide than any other gas except steam. The solution of limestone by waters charged with this gas and its deposition at the surface of the ground where the gas escapes are facts known to every geologist. Jurassic limestones crop out sparingly on the north and south sides of the warm ground at Mammoth and more abundantly at Snow Pass. The travertine in this area varies in "density" according to the rate of deposition. About 9 inches (23 centimeters) a year is the highest rate found in recent measurements. Most specimens of the deposit collected for

study have proved to be calcite, though aragonite has been identified in a few of them.

It is not yet certain how far algae contribute, by their abstraction of carbon dioxide from the water, to the deposition of travertine. In many places the mass of organisms appears entirely too small to be an important factor.

2. *Roaring Mountain*, about 15 miles (24 kilometers) south of Mammoth on the road to Norris Basin, is an interesting example of a pure sulphate area. Comparatively little of the rain falling on this steep mountain slope penetrates into the ground, and the result is a very small discharge, less than 0.1 cubic foot (0.003 cubic meter) a second from all the streamlets combined. Sulphur may be found in many places. Its oxidation and that of hydrogen sulphide in the gases gives sulphuric acid, which, in leaching the rock, carries into solution sulphates of iron, aluminum, calcium, and the alkali metals. Sulphuric acid in the waters reaches a relatively high concentration, 350 to 850 parts H_2SO_4 per million, but the concentration depends quite as much on the rate at which the acid is depleted in leaching the rock as it does upon the rate of oxidation that produced the acid.

Leaching of rock by sulphuric acid is in general a disintegrating process, a reduction to the form of sand and to solutions that escape. At this place, however, silicification or cementation by a part of the opaline silica is also going on.

3. *Norris Basin* is a hot area of mixed character, with a moderately high discharge of water and a relatively large supply of sulphur. Two-thirds of the waters contain mineral matter, much of which is derived from both the deep and the shallow circulations. Only the springs in the northeastern portion of the Porcelain Basin and a few others are alkaline, and they contain far less bicarbonate than other waters of the alkaline type; pure sulphide springs are found only on the borders of the basin around Congress Pool and east of the museum.

Since 1928 the aggregate discharge of the Norris Basin has decreased from 4.35 to 1.35 cubic feet (0.12 to 0.04 cubic meter) a second. This decrease is tentatively attributed to the drought that has prevailed in the park for several years. In the northern portion of the One-hundred Spring Plain many rather shallow springs of mixed origin, but predominantly acid, have shown, summer after summer, marked variations in level as warm weather advanced, variations that have every appearance of seasonal changes.

Several interesting superheated vents are found on the edge of the Porcelain Basin. One of them, the Black Growler, for two successive summers showed a temperature of nearly 138°

C., but it has since varied from 103° to 125° C., an effect attributed directly or indirectly (through choking) to the agency of ground water. The exhalations of the Black Growler in 1928 were 99.6 per cent steam.

In dry weather the Growler Spring, on the north side of the basin, behaves as a superheated vent, varying in temperature from 101° to 104° C., and after heavy rains the crater in which the vent issues partly fills with water and forms a boiling, splashing spring.

In 1926, when this basin was studied, it contained 35 geysers of differing magnitude. Most of them show a highly variable behavior. Many after remaining dormant for months, may become fairly regular for a time. The Fissure or New Geyser has perhaps been the most regular of them all. It plays at a rather sharp angle to a distance of 20 to 30 feet (6 to 9 meters) and has in the past played every few minutes. The Ledge or Mud Geyser in recent years has played infrequently to a distance of 50 feet (15 meters) or more, and the same is true of the Valentine. The Constant is one of the most irregular geysers.

Certain geysers in the Norris Basin are unique in playing directly from narrow fissures in the rhyolite rock instead of from sinter-lined tubes or wells. Such are the Fissure, Ledge, Minute, and Monarch (extinct) Geysers.

Norris Basin represents the newest outbreak of hydrothermal activity in the park, but the thermal energy is exceedingly variable. Within the last 50 years hot springs and geysers have disappeared so completely that they can not be traced to-day, while others have sprung into being. Congress Pool, named in 1891 in honor of the visit of the Fifth International Geological Congress, was then a geyser of considerable size, but it ceased to play years ago.

The remarkable variation in character in the belt of ground extending southward from Congress Pool is due to absence of sinter to cement the superficial material into a compact mass. The frequent changes in the water from clear to muddy, the choking of old vents, and the breaking out of new passages for steam are probably due to shifting of the loose sand and clay. In the summer of 1930 two new superheated vents broke out in this basin, showing temperatures of 112° and 136° C.

On the edge of the basin near Congress Pool, where no thermal activity was visible, the Carnegie Institution in 1930 drilled a 2-inch (5-centimeter) test hole to a depth of 265 feet (81 meters). Steam rushed from the hole at high velocity. When the valve was closed, the pressure inside rose to 300 pounds to the square inch (21 kilograms to the square centimeter), and a temperature

determination at the bottom of the hole gave 205° C. The true temperature was probably higher, because saturated steam at 200° C. has a pressure less than 300 pounds. That the steam escaping from this point was saturated rather than superheated was attributed to the local presence of an excess of ground water.

Deposits from mixed waters in the Norris Basin are usually those characteristic of acid areas—quartz sand covering the ground and a mud consisting of opal, kaolinite, and sometimes sulphur and other minerals in the springs.

Experiments on the deposition of silica proved that some geyser waters in the Norris Basin deposit no sinter in a period of 10 months—in fact, the water of certain rock cracks has produced no accumulation of sinter during a period of six years. On the other hand, the alkaline waters of the Porcelain Terrace precipitated, within 10 months, a layer of opal having a maximum thickness of 32 millimeters. Some of these waters contain a maximum of 0.717 gram of silica per liter—more than any other natural waters known.

The low ridge that marks the western boundary of the Porcelain Basin is a range of the Ragged Hills, consisting of old gravel, probably glacial, leached and cemented by opal. Activity here is now extinct, but a similar deposit about certain springs of the One-hundred Spring Plain is forming to-day.

4. The *Terrace Springs* are chiefly interesting because they deposit both silica and calcium carbonate, partly in mixtures, partly separate. Old terraces on the south side of the road are travertine. The outlet streams have cut their way to slightly lower levels, leaving the terraces entirely dry.

Two other localities, Firehole Lake and Hillside Springs, have characteristics similar to Terrace Springs.

5. The *Fountain Group*, one of the numerous hot-spring groups of the Lower Geyser Basin, shows the characteristics of a typical alkaline area—moderately high discharge, very little sulphur in any form, clear alkaline waters, and siliceous sinter deposits. There is a considerable number of superheated waters and geysers. Clepsydra, Jet, and other geysers play rather frequently to maximum heights of 10 to 25 feet (3 to 7.6 meters). Fountain Geyser, the most powerful of them all, rising 50 feet (15 meters), now seldom plays. On the plain a few hundred yards to the north, the Gentian, a large hot pool more than 40 feet (12 meters) deep, displays a remarkably deep blue color. The color is doubtless due, as Bunsen concluded of geyser waters in Iceland, to a selective absorption of the red light rays and a reflection of the blue. The conditions most

favorable to the blue color are clearness and depth of water and white spring walls and bottom.

The Fountain Paint Pots, near by, exemplify the highly local character of the ground-water supply. Mud pots are not springs reduced by evaporation, but are hot springs in which very slight water supply produces a sulphate water. The mud is a mixture of silica and kaolinite, like that of acid springs.

6. *Excelsior or Midway Basin*, a typical alkaline area in essential characteristics, exhibits two features of special interest. Excelsior Cauldron, 330 feet (100 meters) across in its longest diameter, discharges approximately 6 cubic feet (0.16 cubic meter) of water a second. Once the greatest geyser in the park, it ceased to play more than 40 years ago. It is said at one time to have erupted as often as once an hour, throwing up great masses of water and rocks to a maximum height of 250 to 300 feet (76 to 92 meters). Its violence probably contributed to its own destruction by damaging the underground reservoir in which geyser action is thought to originate.

Prismatic Lake, south of Excelsior, the largest hot pool in the park, is 350 feet (107 meters) in its longest diameter. In 1930 it had an overflow of more than 1 cubic foot (0.03 cubic meter) a second. The temperature, 63° C., is very high for a pool of this size. The blue color of the water in its deeper portions blends on the borders of the pool to a green with the yellow of a peculiar variety of algae, and on the gently descending terraces surrounding the pool a thin growth of a red organism glows with peculiar brilliancy when viewed in bright sunlight.

7. The *Upper Basin*, now representing the climax of thermal activity in the park, contains more deep hot pools, more superheated waters, and more numerous and powerful geysers than any other hot area. Its picturesque features include a number of beautiful hot pools and many remarkable silica mounds, in addition to the world's greatest geysers. A boring in this basin encountered siliceous sinter for the first 20 feet (6 meters), followed by altered rhyolite gravel, probably glacial, for 200 feet (61 meters) and altered rhyolite rock for 200 feet more. This rhyolite gravel, found in all the geyser basins, is probably a factor of some importance in the distribution of ground water. Rock alteration by carbon dioxide and perhaps other gases is not a disintegrator like leaching with sulphuric acid. Much of the material is recrystallized in the form of new minerals, and at the surface of the ground cementation of silica is in progress.

The total discharge of hot water from this basin is about 18 cubic feet (0.5 cubic meter) a second. The variation is probably small. The average temperature is high, a large proportion of the springs surpassing 90° C.

A long series of determinations of winter temperature and barometric measurements show that the temperatures vary, in some springs from a fraction of a degree to 2° C.; in others 5° to 10° or more; that the temperatures do not follow the barometric pressure; and that in a few springs a seasonal effect is unquestionable, the springs being cooled by excess of ground water in winter or spring, but in the greater number of springs the variations can not be ascribed to seasonal differences.

Of all the Yellowstone geysers, Old Faithful is the most regular in its habit and perhaps the most beautiful. Another regular geyser whose jet reaches a maximum length of 75 feet (23 meters) or more (it plays at an angle) is the Daisy.

In the Upper Basin are found numerous hot springs whose surface layer measures from a fraction of a degree to several degrees centigrade above the boiling point for the altitude. The phenomenon of superheating is undoubtedly due to circulation, which causes the deeper and hotter layers of water to rise to the top. These layers are of course in an unstable state and burst into violent boiling when disturbed. In this process the temperature falls, but not to the boiling point, probably because the heat is continually supplied from below. More than a hundred such springs have been discovered in the park.

A study of the deposition of silica in the Upper Basin has shown it to be very irregular, differing greatly at different times as well as in different places. Under natural conditions it proceeds very slowly, from zero to 0.1 inch (2.7 millimeters) a year. In many situations evaporation is undoubtedly the important factor. Where evaporation has been accelerated by artificial means, a deposit of 4.19 feet (1.262 meters) has been obtained in a year's time. Silica can be separated from a geyser water by freezing, and a soft pasty opal has been repeatedly noticed in the field on terraces and along outlet channels after severe winter weather. How much of this soft deposit becomes permanently cemented to the terrace has not yet been determined.

8. Several areas along the shore of the West Thumb of Yellowstone Lake show the same general characteristics as other geyser basins—numerous hot pools, geysers, a few superheated springs, and waters, practically all alkaline, depositing siliceous sinter. On the edge of the basin is a group of mud pots, their water supply even scantier than that of the Fountain Paint Pots. As a result, a thicker mud is found in certain of these pots, clods of which, ejected by escaping steam, build up small cones. The muds of these and other pots show various colors—white, gray, black, and different shades of red. As the essential constituents of the mud are always opal and kaolinite, the color is always

white except when modified by small amounts of other minerals. The reds are due to varying amounts of hematite; the grays and blacks to finely divided pyrite.

9. *Mud volcano group.*—The hot muddy pools along the Yellowstone River 6 miles south of Lake Junction are nearly pure sulphate waters. The largest pool at the south end was once the seat of a true "mud geyser," said to have played to a height of 20 to 30 feet (6 to 9 meters). Down the road 0.2 mile (0.3 kilometer) from this point, at the foot of a high steep bank, is an acid area, devoid of any discharge, containing what is probably the largest and certainly one of the most active mud pots in the park.

BIBLIOGRAPHY OF YELLOWSTONE PARK

1. Holmes, W. H., Report on the geology of the Yellowstone National Park: U. S. Geol. and Geog. Survey Terr. Twelfth Ann. Rept., pt. 2, pp. 1–62, 1883.

2. Iddings, J. P., Obsidian Cliff, Yellowstone National Park: U. S. Geol. Survey Seventh Ann. Rept., pp. 249–295, 1888.

3. Hague, Arnold, and others, U. S. Geol. Survey Geol. Atlas, Yellowstone National Park folio (No. 30), 1896.

4. Iddings, J. P., The igneous rocks of Electric Peak and Sepulchre Mountain, Yellowstone National Park: U. S. Geol. Survey Mon. 32, pt. 2, pp. 89–148, 1899.

5. Goode, J. P., The piracy of the Yellowstone: Jour. Geology, vol. 7, pp. 261–271, 1899.

6. Martonne, Emmanuel de, Le Parc National du Yellowstone; esquisse morphologique: Am. Geog. Soc. Memorial Vol., Transcontinental Excursion, 1912, pp. 231–250, 1915.

7. Day, A. L., and others, The temperatures of hot springs and the sources of their heat and water supply [symposium]: Jour. Geology, vol. 32, pp. 177–225, 291–310, 373–399, 499–471, 1924.

8. Foshag, W. F., The minerals of Obsidian Cliff, Yellowstone National Park, and their origin: U. S. Nat. Mus. Proc., vol. 68, art. 17, 18 pp., 1926.

9. Jones, O. T., and Field, R. M., The resurrection of the Grand Canyon of the Yellowstone: Am. Jour. Sci., 5th ser., vol. 17, pp. 260–278, 1929.

10. Jones, O. T., In the Yellowstone with Princeton: Nature, June 1, 1929.

11. Jones, O. T., History of the Grand Canyon, Yellowstone National Park, Royal Inst. Great Britain, April 19, 1929.

12. Fenner, C. N., and Day, A. L., Bore-hole investigations in the geyser basin of Yellowstone Park: Washington Acad. Sci. Jour., vol. 21, No. 20, pp. 488–491, 1931.

MAMMOTH HOT SPRINGS TO GRAND CANYON OF THE YELLOWSTONE

By R. M. Field

This section illustrates the general petrology of the rhyolite plateau and the principal geyser basins. Particular attention will be paid to the lithology and mineralogy of Obsidian Cliff and the features characteristic of the Norris and other geyser basins.

Mammoth Hot Springs.—Ascent of terrace and study of hot springs and spring deposits. From the summit of the terrace to

the east is seen the valley of Lava Creek as far as Undine Falls. The escarpment over which Lava Creek falls is the remnant of the oldest postrhyolite or valley flow (basalt). Beyond Undine Falls Lava Creek turns sharply to the south; the relatively dry upland valley that continues to the east and southeast is the old glacial-overflow channel, which may be traced, with numerous ramifications, as far as Tower Falls. Where the valley widens out below Undine Falls there is no trace of basalt or basalt scree on the south valley wall of Mount Everts. Mount Everts is composed of Mesozoic formations, capped with a lava flow.

Golden Gate (5 miles, or 8 kilometers).—Just before reaching the canyon and rapids called Golden Gate the road passes through a relatively recent landslide of terrace-spring deposits. On the west wall of the canyon may be seen the remnant of the original terrace. The rhyolite is well exposed in the west wall of the canyon just beyond the landslide.

Obsidian Cliff (12 miles, or 19 kilometers).—Wayside Museum on west of road and Obsidian Cliff just beyond on east side of road.

The cliff presents a partial section of a late flow of obsidian which poured down an ancient slope of rhyolite from the plateau lying to the east. * * * The exact point at which this obsidian broke through the older rocks and reached the surface has not yet been discovered; but that forming Obsidian Cliff has evidently flowed down from the high plateau in a northwest direction into a preexisting valley, the planes of flow in the lava clearly indicating that it has crept down the slope back of Obsidian Cliff and accumulated in the bottom of a channel between rhyolite hills.

The obsidian of Obsidian Cliff is a dark glassy rock with conchoidal fracture * * * and usually free of phenocrysts, though the rhyolites elsewhere in the park show well-formed crystals of feldspar scattered through a glassy groundmass. The rock presents all gradations from that purely hyaline to a lithoidite, with but a minor amount of glass. One of the features to first attract attention is the abundance of stony spherulites scattered through the glass, sometimes singly, sometimes aggregated into bunches, and often coalesced to form well-defined layers parallel to the line of flow. In some specimens the glass carries hollow spherulites lined with a white coating of crystalline material, either sparsely disseminated or so abundant as to make up the bulk of the rock. * * * Through an increase in the proportion of stony structure there is a complete gradation of the glassy obsidian into a rock entirely crystalline, though usually microcrystalline—the lithoidite (soda rhyolite composed of tridymite and soda sanidine). * * * Both the obsidian and the lithoidite carry abundant hollow cavities lined with crystalline matter. * * * It is apparent that the layers of lithophysae, like the simpler forms, are intimately connected with the stony spherulites, since they are usually alined with them and in many cases appear to be but special instances of spherulitic crystallization. The lithoidite, spherulites, and lithophysae all have essentially the same composition. The lithophysae attain their greatest complexity and beauty in the completely lithoidal rock. * * * The minerals identified in the cavities and lithophysae of the Obsidian Cliff lavas are, in order of their abundance, feldspar, tridymite, cristobalite (in small "pellets"), fayalite, and quartz; numerous minute needles associated with the feldspars may be hornblendes (2, p. 255).

Norris Geyser Basin (5 miles, or 8 kilometers).—Inspection of the museum illustrating the geology of the basin and the hot springs. Inspection of the hot springs. This basin is particularly remarkable for emanations and minerals. Among those which may be observed are sulphur, arsenic in unknown crystal form (chemical composition $H_3As_2O_4$), sulphides of arsenic (As_2S_3) chlorine in unknown form, alunite, pyrolusite, barium sulphate, bicarbonates and carbonates of sodium and potassium ($NaHCO_3$, Na_2CO_3, $KHCO_3$, K_2CO_3), sodium chloride and potassium chloride, geyserite (SiO_2), pyrite (FeS_2), scorodite ($FeAsO_4.H_2O$), and alunogen ($Al_2O_3.3SO_3.16H_2O$). Scorodite is leek-green in color when fresh but rapidly alters to limonite upon exposure. Alunogen is the fibrous deposit found close to the steam vents and is the result of the oxidation of the sulphates.

Madison Junction (14 miles, or 23 kilometers).—Inspection of museum illustrating the history of the exploration of the Yellowstone Park. A quarter of a mile farther up the canyon is an excellent exposure illustrating the manner in which the rhyolite flowed and cooled.

Lower Geyser Basin (7 miles, or 11 kilometers).—View of Paint Pot. Boiling white mud.

Upper Geyser Basin (10 miles, or 16 kilometers).—Old Faithful Geyser and springs. Sponge Geyser. Old Faithful is remarkable both for its regularity (every 60 to 80 minutes) and the height of its eruption (120 to 170 feet (37 to 52 meters) for four minutes).

Continental Divide (6 miles, or 9.6 kilometers).—Altitude 8,262 feet (2,518 meters). Lake Isa delivers water to both the Atlantic and the Pacific. About 7 miles (11.2 kilometers) farther on the Continental Divide is crossed again at an altitude of 8,362 feet (2,549 meters).

View of Yellowstone Lake (2½ miles, or 4 kilometers).—From this point the road descends rapidly to the ranger station at West Thumb, and the journey continues close to the west shore of Yellowstone Lake, crossing excellent examples of bay-head bars over which the road has been built. The former level of the lake is shown by two benches, which are more clearly exposed near the outlet of the lake at Lake Junction (20 miles, or 32 kilometers).

Dragon's Mouth and Hayden Valley (6 miles, or 9.6 kilometers).—Hot spring rising through till. A mile (1.6 kilometers) farther on is a view of the glacial moraines that dammed the glacial Lake Yellowstone to a sufficient height to divert the drainage, for a time, across the Continental Divide. The moraines have been breached by the Yellowstone, with the formation of deltas sloping gradually to the north. Good sections of the peculiar rhyolite till and lake deposits are exposed along the

road. One mile from Dragon's Mouth, on the left side of the road, is the Monad, a miniature meandering stream with cut-offs which have been figured in several American textbooks of geology.

Canyon Camp (6 miles, or 9.6 kilometers).—Cross bridge to east side of canyon. At top of hill, on east side, just before reaching camp, is an exposure of varied glacial clays, highly contorted. Visit Red Rock, showing section of lake sediments, and Artist Point.

GRAND CANYON OF THE YELLOWSTONE TO ROOSEVELT LODGE

By R. M. Field

The region between the Grand Canyon of the Yellowstone and Roosevelt Lodge affords the best opportunity to study the relative order of the volcanic events in the park, together with the peculiar topography which has resulted from the reexcavation of the Oligocene and Pliocene Valleys.

Return to west side of canyon. A mile (1.6 kilometers) from the bridge, on the west side of the curving and ascending road, is an exposure of the rim conglomerate resting upon andesitic breccia. Across the canyon the camp rests on the rim conglomerate, underlain by sediments which, in turn, rest upon the decomposed rhyolite of the canyon wall. A quarter of a mile (0.4 kilometer) farther up the road the rim conglomerate is overlain by till. A mile (1.6 kilometers) farther on is Grand View.

Mount Washburn (19 miles, or 31 kilometers).—Altitude 10,317 feet (3,144 meters). On the winding road up the south side of the mountain, beyond Dunraven Pass, acid lavas are exposed well above the general level of the plateau. Farther up the road are good exposures of the andesitic breccias. From the summit of the mountain, looking south, is obtained a view of the park plateau. To the northwest the surface of the rhyolite rises gradually toward the mountain and fills the valleys between the spurs. At its highest point the rhyolite is several hundred feet above the general level of the plateau.

Descent of Mount Washburn to Tower Falls and Roosevelt Lodge (11 miles, or 17.7 kilometers).

Petrified Forest.—On the south side of a hill half a mile (0.8 kilometer) south of the main road are several petrified trees. Trunks as well as limbs and roots are exposed. These trees are a portion of extensive forests which are exposed at many places on all the surrounding slopes. The fossil trees occur at nine or more horizons, indicating the repeated destruction of

SOUTH FACE OF HEART MOUNTAIN

Courtesy of Northern Pacific Railway. Photograph by Brown, St. Paul. Shows thrust plate of Madison limestone resting on upturned Paleocene and Eocene beds. Terraces and beveled surfaces in foreground.

ELK BASIN OIL AND GAS FIELD

Courtesy of Northern Pacific Railway. Photograph by Brown, St. Paul. Shows part of surface of Polecat Bench. Looking northeast toward abandoned valley at Pryor Gap.

living forests by volcanic ejectamenta. The agglomerates are intercalated with fine-grained white ash beds which contain well-preserved Eocene to early Miocene leaves. The tree trunks belong to the genus *Cupressinoxylon*. Other fossil forests contain sequoias, 2 species of pine, 2 of oak, 1 of laurel, and other forms. The fossil leaves have been identified as belonging to more than 150 species and have been regarded as most nearly allied to the Fort Union flora of the lower Yellowstone Valley, in eastern Montana.

Junction Butte and Tower Falls.—The road from Camp Roosevelt to Tower Falls ascends a narrow valley (glacial overflow channel) which, near the top of the rise, is bounded on the canyon side by a high wall of basalt with a band of marked columnar structure at the base. The upper portion also is columnar, but the columns are smaller, inclined in various directions, and arranged irregularly. Beyond this mass (1 mile (1.6 kilometers) from the lodge), where the road becomes level, there is a recently drained flat. After crossing this open flat and following the trail to the rim of the canyon a good view is obtained of the opposite wall. On the left the opposite rim of the canyon is formed by a markedly columnar basalt about 25 feet (7.6 meters) thick. Below the basalt is a conglomerate of well-rolled pebbles, about 100 feet (30 meters) thick. Beneath the conglomerate is another band of similar columnar basalt. The lower part of the section is obscured by conglomerate scree. The upper basalt can be seen for a considerable distance up the canyon and maintains its characteristics and thickness in that direction. The conglomerate, however, diminishes to less than half its thickness as it is traced up the canyon, and the lower basalt disappears. In that direction the lower part of the conglomerate and the lower basalt are replaced horizontally by a poorly stratified conglomeratic rock (andesitic breccia), which extends to the bottom of the canyon.

North of Tower Falls, at Hanging Cliff, the east wall of the canyon shows the upper basalt to be underlain by the conglomerate to the bottom of the valley. Thus the lower basalt and the lower beds of the conglomerate pass behind the wall of andesitic breccia. This wall is a screen between the modern canyon and the older canyon, which is concealed behind it. Hanging Cliff consists of a sheet of basalt, with a well-defined band of large columns near the base and small irregular columns in the upper part. On descending the hill toward Tower Falls, the basalt is seen to rest upon a few feet of conglomerate, with a few inches of muddy silt between. Farther on the basalt shows pillow structure and other evidence of having flowed into water. This basalt and the underlying conglomerate are the same beds

that form the upper part of the canyon on the east side. The difference in the thickness of the conglomerate is due to the fact that the surface on which it was deposited lies at a much higher level on the west side than on the east side. The gravel was laid down against the flank of a valley. On looking north from Tower Creek, the basalt can be seen banked up against the same valley slope at a higher level. The relations of the rocks on the east and west side of the canyon are diagrammatically expressed in Figure 4.

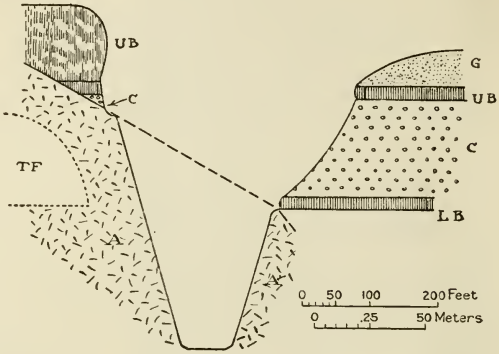

Figure 4.—Diagrammatic section across Yellowstone Canyon near Tower Falls. A, Andesitic breccia; LB, lower basalt; C, canyon conglomerate; UB, upper basalt; G, glacial deposits; TF, lower basalt near Tower Falls projected to line of section. The dashed line shows portion of former valley floor

ROOSEVELT LODGE TO EAST ENTRANCE OF YELLOWSTONE PARK

By R. M. Field

Return to Lake Junction and thence through Sylvan Pass to east park entrance. Inspection of museum at Fishing Bridge Camp (1 mile (1.6 kilometers) from junction). Continuing through Sylvan Pass (21 miles, or 34 kilometers) are excellent exposures of the andesitic breccia. About 3 miles (4.8 kilometers) farther on basalt dikes cut the andesitic breccia.

EAST ENTRANCE OF YELLOWSTONE PARK TO CODY

By W. T. Thom, Jr., R. M. Field, R. T. Chamberlin, and W. H. Bucher

On the road to Cody the descent is made first through volcanic rocks, and then through beds ranging from Eocene to Cambrian in age. Partial cross sections can also be seen of the great lime-

stone plate beneath the volcanic rocks, which has been overthrust eastward above the Rattlesnake Mountain anticline for a total distance of 18 miles (29 kilometers) or more.

Rattlesnake Mountain is an asymmetric fold such as is characteristic of the eastern Rocky Mountains, and the Shoshone Canyon affords clear evidence as to the relation of the fault in the granite basement to the sharp flexure in the sediments above. The course of the Shoshone River across this anticline is an excellent example of stream superposition, characteristic of Wyoming.

Beyond the east entrance to Yellowstone Park the road follows first Sylvan Pass Creek and then the valley of the North Fork of the Shoshone River, and the traveler, passing thus eastward, gradually descends between towering cliffs and ridges, finally reaching the base of the volcanic series near the eastern boundary of the national forest, about 30 miles (48 kilometers) east of the park entrance. Dikes, flows, and deposits formed of water-borne volcanic materials are abundantly exposed. The vertical jointing and unequal hardness of the rocks have resulted in the production of many pinnacles of fantastic shapes.

At Chimney Rock, 12 miles (19 kilometers) from the east entrance, is a good exposure of agglomerates and sandy beds containing plant remains.

At a point 29½ miles (47.4 kilometers) east of the park entrance tawny sandstones and varicolored clays of Eocene age appear from beneath the volcanic deposits, and 2 miles (3.2 kilometers) farther on the Tertiary beds are cut by a large dike, which is visible on both sides of the highway. In this vicinity terraces mantled with gravel are conspicuously developed along the valley, and at the bridge 33 miles (53 kilometers) from the park entrance the river flows in a narrow trench cut in a broad gravel-covered plain.

At 2½ miles (4 kilometers) east of the bridge an abrupt flexure or fault makes the eastward termination of the Tertiary. Here also first evidences become visible of the great thrust faults of prevolcanic age, which border the park block on the east, the Tertiary beds being here overlain by remnants of a great thrust plate of Ordovician (Bighorn) limestone which was pushed eastward for 18 miles (29 kilometers) or more, overriding, as it moved, the Rattlesnake Mountain anticline, which now becomes visible as a wall almost blocking the valley toward the east. (See fig. 5.)

Slightly more than a mile farther east the last remnants of the volcanic rocks are seen on the high ridge north of the highway, filling deep channels eroded in the Madison limestone thrust plate.

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At 39 miles (63 kilometers) from the park entrance the road comes abreast of the west end of the Shoshone Reservoir. South of the lake a part of the limestone thrust sheet forms the towering cliffs of Sheep Mountain, rising above lower slopes of Cretaceous shales and sandstones, which likewise are exposed in the road cuts.

About 3 miles (4.8 kilometers) farther on, at the bridge over Rattlesnake Creek, the edge of the overthrust limestone mass is visible to the north as the escarpment of Logan and Chalk Mountains, and it can be seen converging toward the Rattlesnake Mountain fold, east of which a more advanced remnant of the thrust sheet persists as the cap of Heart Mountain, a conspicuous landmark of the western Big Horn Basin. From this

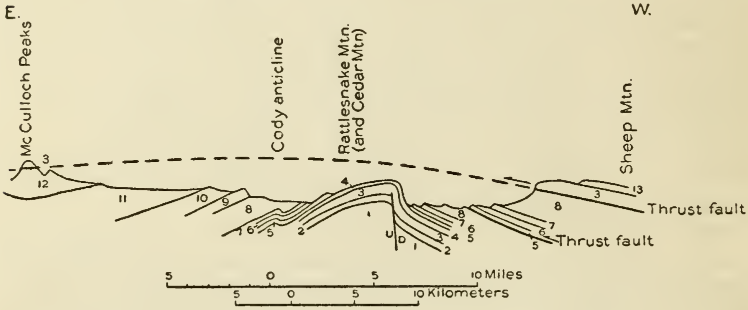

FIGURE 5.—Structural cross section from Sheep Mountain to McCulloch Peaks. 1, Granite; 2, Cambrian; 3, Ordovician and Mississippian; 4, Pennsylvanian and Permian; 5, Triassic; 6, Jurassic; 7, Lower Cretaceous and Dakota; 8, Upper Cretaceous (lower part); 9, Upper Cretaceous (Mesaverde);. 10, Upper Cretaceous (Lance); 11, Paleocene (Fort Union); 12, Eocene (Wasatch and Bridger); 13, volcanic rocks

point, too, are clearly seen the vertical Paleozoic beds rising as a sort of palisade along the west side of Rattlesnake Mountain.

The visitor is also impressed by the anomalous course of the Shoshone River, which, instead of following the shale valley that encircles the plunging south end of the Rattlesnake Mountain anticline, has cut a great canyon across the fold about 4 miles (6.4 kilometers) from its southern termination, thereby permitting the creation of a great storage reservoir by the construction of a relatively narrow dam in the throat of the gorge. Presumably the position of the Shoshone Canyon has been determined by superimposition, either directly on the thrust plate or on overlying sediments or volcanic rocks.

Southeast of the lake strike ridges of Cretaceous and Jurassic sandstones project above the blanket of terrace deposits, and

the Chugwater red beds (Triassic) are exposed in places along the foot of the mountain slopes.

The Embar limestone (Permian) and Tensleep sandstone (Pennsylvanian) together form the outer ridge or dip slope of the mountain and are underlain by the less resistant reddish shales and varicolored limestones of the Amsden formation, which tends to erode into strike valleys between the outer ridge of Embar and Tensleep and the inner, stripped surface of the Madison limestone. The Madison and the underlying Bighorn dolomite (Ordovician) form a very resistant unit, which is underlain by Cambrian limestone (Gallatin) and soft greenish and plum-colored glauconitic shales (Gros Ventre).

Near the Shoshone Dam the road passes through two tunnels cut through Pleistocene and Recent calcareous hot-spring deposits in which some fossil leaves are embedded; and at the dam a third tunnel is cut through the pre-Cambrian granite along the plane of a thin horizontal sheet of basic rock which is clearly exposed in the tunnel walls.

From exposures at the dam and others afforded by the canyon farther east, it is evident that the Rattlesnake Mountain anticline is a strongly asymmetric fold, with vertical or nearly vertical dips on the west limb and only moderate dips on the east limb. At the dam it can also be seen that the granite on the east has been faulted up relative to the Cambrian and Ordovician shales and limestones on the west but that this fault dies out upward, merging into a sharp monoclinal flexure, such as is characteristic of eastern Rocky Mountain anticlines.

East of the dam the road is cut in dark biotite granite, much shattered and injected by pinkish pegmatitic dikes and sheets.

At 0.6 mile (0.96 kilometer) below the dam, or 47 miles (76 kilometers) from the east park entrance, a large basic dike, which has come up along a fault of small displacement with downthrow to the south, can be seen at the bottom of a deep notch in the northwest canyon wall, its strike being directly down the canyon toward the east.

A little farther east the granite passes beneath the basal Cambrian arkose or quartzite, and the stratigraphic section seen above the dam is then retraversed, in ascending order.

Within the lower canyon deposits left by extinct hot springs are numerous, and terrace deposits, representative of at least two of the terrace stages developed above the canyon, are also preserved along the canyon sides. Just south of the bridge at the lower end of the canyon is the Shoshone Cavern National Monument. As the east portal of the gorge is approached the traveler is impressed by the number and perfection of the gravel-covered terraces which border the lower Shoshone Valley.

The Chugwater red beds can be seen downstream (near the DeMaris Hot Springs), and also to the north, where they rise in a gentle anticlinal arch along the mountain side. Lesser ridges formed by the Sundance formation (Jurassic) and by the Morrison and Cloverly sandstones intervene between the "Red Beds" arch and the conspicuous peaks of Heart Mountain, which are capped by Madison limestone, a remnant of the overthrust plate of Paleozoic limestone, here resting on vertical Fort Union and strongly tilted Wasatch beds.

South of the highway, beyond the canyon, are two hot springs now actively depositing sulphur (hence Stinking Water River, the old name of the Shoshone River), which is being mined in a small way, and about a mile (1.6 kilometers) farther on exposures north of the road reveal dark Cretaceous shales dipping eastward at moderate angles.

At 53 miles (85 kilometers) from the park boundary is the Buffalo Bill Museum, at the entrance to Cody, near which an equestrian statue has been erected in honor of Colonel Cody.

CODY TO RED LODGE

By W. T. Thom, Jr., R. T. Chamberlin, W. H. Bucher, W. J. Sinclair, and G. L. Jepsen

Between Cody and Red Lodge the route has been so chosen as to afford opportunities for study of the sedimentary formations and the physiographic and structural features typical of Rocky Mountain intermontane basins. The area crossed is, furthermore, one in which an unusual succession of primitive mammalian faunas has been obtained (see pp. 35–36 and Appendix A) and one which illustrates in a measure the part played by petroleum development in bringing about the exploration and settlement of Wyoming (see Appendix B) and other Western States.

The topographic and structural features of the region around Cody are shown by Plate 4 and Figures 5 and 6.

At 7 miles (11 kilometers) from Cody, near the bridge, is an exposure of dipping basal Wasatch beds truncated and covered by a thin veneer of gravel consisting in considerable part of volcanic rocks. The relationship of beveled beds underlying the present thin gravel mantle indicates clearly the character of this and other surfaces of stream planation and subaerial denudation, which are characteristic of this region.

About 9 miles (14 kilometers) from Cody is a viewpoint for the McCulloch Peaks, which lie in a syncline near the axis of the major basin trough. These peaks are composed of Wasatch and Wind River (Eocene) beds, capped on the farthest summit,

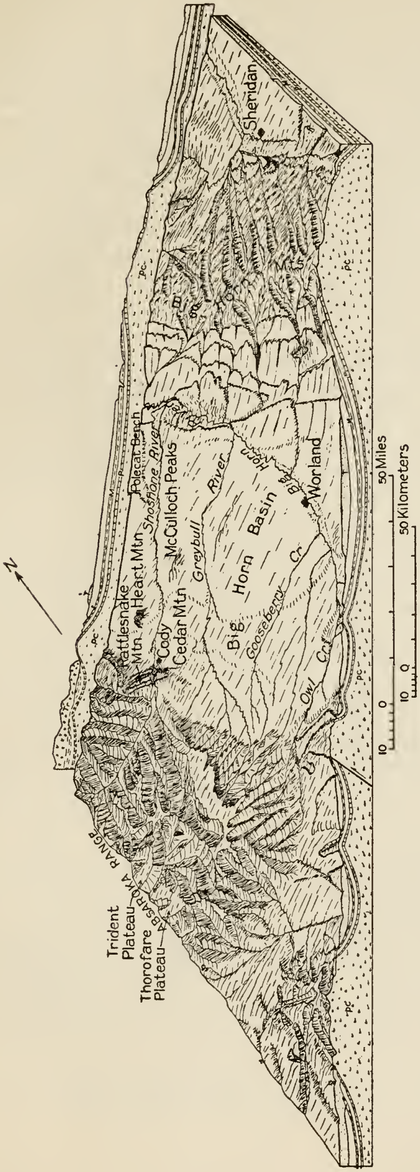

FIGURE 6.—Block diagram showing Absaroka Mountains, Big Horn Basin, and Big Horn Mountains

which is not seen from this point, by Ordovician and Mississippian limestone, an advanced remnant of the Heart Mountain overthrust sheet. Certain of the Wasatch beds are tinted pink, the coloring appearing more striking on weathering.

Polecat Bench,[3] at Ralston, 19 miles (31 kilometers) from Cody, is a great level-topped topographic feature, cut across the edges of southwestward-dipping strata. This erosion remnant, surrounded on all sides by a steep escarpment, is shown on the geologic map of Wyoming as covered by Quaternary gravel. (See also fig. 7.) Its truncated dipping strata include the Wasatch, Fort Union, Lance, and underlying Cretaceous beds, which are veneered by soil and by local gravel beds. To the east are seen the Big Horn Mountains and to the northeast the Pryor Mountains.

At 3 miles (4.8 kilometers) north of Powell can be seen the beveling of the Wasatch and Fort Union beds, which dip toward the west. These strata have furnished very significant vertebrate fossils and afford the most complete stratigraphic section yielding archaic mammals yet discovered in North America. (See Appendix A.)

The top of the Polecat Bench escarpment is 33 miles (53 kilometers) from Cody. Volcanic rocks are most abundant in the gravel here, apparently having been transported from the areas of Tertiary volcanic rocks in the Absaroka Mountains to the west. A longitudinal central valley on the bench indicates the nature of the relief before rejuvenation, and the grade and orientation of the valley indicate that the Shoshone River once flowed across Polecat Bench at the time that the river was occupying the upper, open part of Shoshone Canyon. A ridge of Fort Union beds extending south from Heart Mountain apparently was truncated by the river when it was flowing on this grade.

To the northeast Polecat Bench appears also to correspond in altitude with the Pryor wind gap, apparently indicating that the Shoshone River once passed through the Pryor gorge. Later the more powerful Big Horn River farther east is thought to have tapped the Shoshone and carried its water around the east side of the Pryor Mountains, perhaps after Clark Fork had been beheaded by a tributary of the Yellowstone which cut headward along the strike.

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[3] The term "bench" is used in these descriptions to signify a nearly flat-topped remnant of a surface of low relief cut across dipping beds.

To the south the observer looks down upon the broad flats of the present Shoshone, cut across the edges of tilted beds, deeply soil covered and under cultivation as a part of the irrigation project supplied by the Shoshone Reservoir west of Cody. Scarcely more than 20 years ago these broad terrace flats were practically uninhabited, treeless, with a scanty growth of sagebrush, cactus, and other desert plants, and much frequented by rattlesnakes. The present cultivation is a fair sample of what can be accomplished by the irrigation of desert lands.

Where the motor road reaches the top of Polecat Bench a wagon track leads from the main road northeastward to an amphitheater eaten into the margin of the bench by the headwaters of Polecat Creek. Here beds of the Montana group, the Lance, and the Fort Union are involved in local anticlinal and synclinal folds and are faulted, giving rise to hogbacks intersected and dropped by the faulting. The stratigraphic sequence here exposed is shown in Figure 7. The view may be correlated with the diagram by means of the cliff-forming yellow-brown sandstone, about 130 feet (40 meters) thick, which is seen dipping southwestward into the cliff face below the observer. The somber-colored shales with thin sandstone layers above this sandstone cliff carry a few coal beds, which have given rise to several abortive mining ventures. About 15 feet (4.5 meters) above the highest of these coal beds lies a sandstone, 9 to 31 feet (2.7 to 9.4 meters) thick, which at the base is white with many flattened blue clay galls and grades upward into a brownish phase with few clay galls. This sandstone carries fish scales, reptile teeth and skeletal parts, and turtle bones. Its basal whitish facies has yielded a small mammalian fauna of Torrejon (Paleocene) age. The 130-foot (40-meter) cliff-forming sandstone previously mentioned rests on a thin bed of coal. Impressions of compressed cycadlike trunks may be seen here and there in the lower surface of this sandstone. Locally the coal has been eroded away, and basal layers of the sandstone abound in clay galls. Here a small mammalian fauna of Puerco (early Paleocene) age was found. In the shales 20 feet (6 meters) below the base of the sandstone were found dinosaur bones which prove them to be of Lance age (here regarded as uppermost Cretaceous). (For a general account of the stratigraphy and mammal horizons of this classic region see Appendix A.)

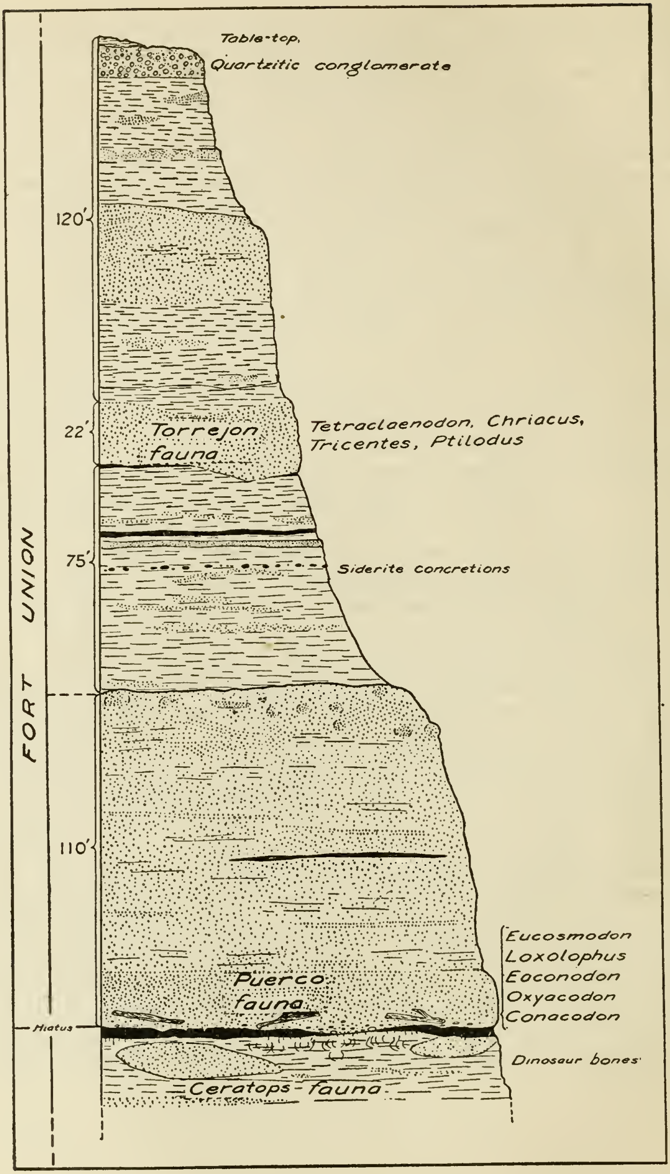

FIGURE 7.—Generalized stratigraphic section exposed in Polecat Bench escarpment, sec. 31, T. 57 N., R. 98 W. (After Jepsen, Am. Philos. Soc. Proc., vol. 69, p. 479, 1930)

The Elk Basin anticline and oil field (see pl. 5 and fig. 8) is 42 miles (68 kilometers) from Cody. The central portion of the basin exposes the upper portion of the Colorado shales. The inner basin rim, seen along the road, is made of a lower gray sandstone (Elk Basin) and an upper yellowish sandstone (Eagle), in the upper part of which is a coal bed. These beds continue the rim to the far end of the basin. To the west and north is a distant outer rim composed of Montana and Lance formations. Eight or ten large faults cut across the basin transverse to its long axis. (See fig. 8.) Their planes are

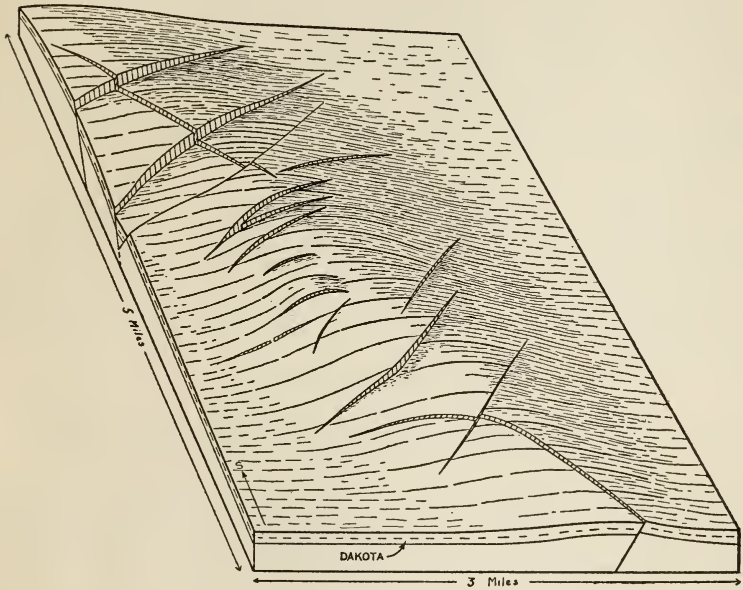

FIGURE 8.—Block diagram showing structure of Dakota (?) sandstone in Elk Basin field. (By A. J. Eardley)

steep and dip in opposite directions, and some of them define wedges. The Polecat Bench bevels the entire Elk Basin structure, and the basin has since been excavated by renewed differential erosion.

To the west is a striking view of the subsummit peneplain on the granite core of the Beartooth Mountains, and to the northwest is the Mesa Bench surface, which bevels Fort Union and probably Wasatch beds. Heart Mountain lies to the southwest and the McCulloch Peaks to the south. The Big Horn Mountains lie to the east and the Pryor Mountains to the northeast.

Elk Basin yielded the first oil obtained in Montana and has been producing oil and gas for nearly 20 years. (See Appendix B.)

Along the road leading out of Elk Basin may be seen the Elk Basin sandstone, constituting the inner rim of the basin; the Eagle sandstone, with coal in the upper part, forming a second rim; and the Parkman sandstone, making a third rim. The road turns north in a valley of estuarine beds at the Bearpaw horizon (topmost Upper Cretaceous).

Below Belfry the Hell Creek member of the Lance formation appears in the river bluff, and between Belfry and Bear Creek the road traverses the Tullock member of the Lance and the Lebo and Tongue members of the Fort Union. Two miles (3.2 kilometers) southwest of the town of Bear Creek, up Foster Gulch, is the Eagle coal mine, which has yielded a large mammalian and reptilian fauna, and closely associated with this fauna stratigraphically are beds carrying a large fossil flora. (See p. 42.)

At the top of Bear Creek Hill, 75 miles (121 kilometers) from Cody, there is a view of the near-by vertical and locally overturned Paleozoic formations (chiefly Madison and Bighorn limestones) in the Beartooth front and of the Mesa Bench, sloping away from the mountains. Along the road leading downhill to Red Lodge twin basic dikes are seen cutting the Fort Union.

RED LODGE AREA

By N. M. Fenneman, Erling Dorf, W. J. Sinclair, and G. L. Jepsen

[Alternative with excursion Red Lodge to Beartooth Butte]

The area surrounding Red Lodge affords many features of highly varied geologic interest. Exposures in the upturned mountain front and adjoining badlands reveal several thousand feet of Paleozoic, Paleocene, and early Eocene strata, and bench-gravel deposits of Tertiary(?) and Quaternary age cover large areas. Thrust-faulting of the great anticline, which has given rise to a total differential uplift of the mountains of at least 5 miles (8 kilometers), is clearly shown by the form of the mountain front. The local land forms bear witness to the fact that the area has had a long and varied physiographic history; and the marine faunas of the Paleozoic formations, the fossil floras of the Paleocene, and the associate vertebrate and invertebrate faunas lend the locality unusual interest to the paleontologist.

Bedrock beneath Red Lodge consists of Fort Union coal measures, and the branch line of the Northern Pacific Railway that extends to Red Lodge was built to reach the coal mines around which the town has grown. Exposures of the coal measures and

of one of the coal beds are to be seen in the creek bank adjoining the mine near the east edge of town. The local coal measures dip about 18° SW. here but are flatter toward the mountains. They are mantled by bench gravel, as may be seen from a vantage point 2 miles (3.2 kilometers) from town on the old Red Lodge-Cody wagon road.

Here can be seen the present Rock Creek flood plain and the second bench level (Red Lodge terrace), which is well developed between Rock Creek and West Fork. A third bench (the Roberts Bench) lies in the immediate foreground, and at this point is the southwest end of a remnant of the highest or Mesa Bench. These features are related to stages in the cutting of the present canyons, and bench surfaces can be seen on spurs within the canyons, thus permitting a partial restoration of the old stream gradients.

Near Red Lodge the Mesa and Roberts Benches have gradients similar to that of Rock Creek, but at a distance of 10 to 15 miles (16 to 24 kilometers) downstream their profiles flatten and the stream is more deeply intrenched. Such a relation may be expected to follow an uplift with little or no tilt. The second bench, however, seen mainly on the west side of Rock Creek, has a much steeper gradient than the stream and correspondingly steeper than the higher benches, being 100 feet (30 meters) above the flood plain at Red Lodge but steadily declining and merging with the flood plain in about 15 miles (24 kilometers). This suggests an upward movement of the plains.

Aggradation by Rock Creek is recorded in the thick boulder and gravel flood-plain deposit on which the city of Red Lodge stands. This deposit is known to be at least 93 feet (28 meters) deep and may be much deeper. It was probably deposited by overloaded glacial streams from the mountains.

The varying thickness of the coarse torrential deposits coating the Mesa peneplain indicates that this graded surface was partly made by the filling of valleys. Possible crustal movements since the formation of the Mesa peneplain are not yet satisfactorily worked out.

Beyond the lowland benches are the vertical Paleozoic limestones, which form a border to the higher granite ridges that mark the dissected front of the Beartooth Plateau.

The abrupt mountain front corresponds with the eastern front of the overthrust mountain mass and in part owes its present prominence to the relative ease with which erosion has removed the soft Fort Union sediments, where not protected by gravel beds. The gradual migration of the divide away from Clark Fork and toward Rock Creek, as a result of the stronger flow and deeper intrenchment of Clark Fork, is also evident when one

looks at the gullied slopes leading down from the Roberts Bench to Bear Creek. This condition contrasts with the lack of tributary gullies along the corresponding stretch of Rock Creek.

The eastward thrust of the mountain mass has not been uniform, as may be seen by tracing the tilted Paleozoic limestones northward from Grove Creek Point. A nearly east-west fault along the foot of Mount Maurice (figs. 9, 10) displaces the limestones eastward on the south side of the fault about 3 miles (4.8 kilometers). To the northwest the limestones continue in a somewhat curved line to Willow Creek, where, owing to another fault the northern limestone ridge is offset eastward about half a mile (0.8 kilometer).

As the traveler descends the hill and goes toward Grove Creek Point, drag on the Mount Maurice fault can be seen in the Gallatin limestone (Cambrian), which is abruptly flexed and brought within 100 feet (30 meters) of the granite. There is also a valley, parallel to the mountain front, which is cut in the granite instead of in the overlying Cambrian shale (on which the "switchback" road ascends). Apparently this valley was developed along the strike of the Cambrian shales when they extended much higher upon the mountain side than at present.

At Grove Creek Point the Fort Union beds dip gently westward at the foot of the mountain, and within 1 mile (1.6 kilometers) of the front a few hundred feet of coarse conglomeratic sandstone, presumably the basal measures of the Wasatch formation, rests with apparent conformity upon the Fort Union.

At the start of the ascent of the switchback road can be seen the drag on the Mount Maurice fault and also the old wave-cut surface on the pre-Cambrian granite.

The road follows the strike of the Gros Ventre shales, and at the divide at the top of the first ascent on the switchback road the Flathead quartzite crops out between the Gros Ventre and the granite. Overlying the Gros Ventre shales are the Cambrian Gallatin limestones, flat-pebble conglomerates, and shales; Upper Ordovician Bighorn dolomite; Devonian Jefferson limestone; lower Mississippian Madison limestone; red shales and limestone conglomerate probably belonging to the Amsden formation (late Mississippian and early Pennsylvanian); and obscure limestone outcrops perhaps representative of the Phosphoria formation (Permian).

The flexing of the Paleozoic beds is by no means uniform, one underturned fold in the Gallatin limestone being conspicuous, but in general the limestones are somewhat overturned at the eastern base of the mountains and tend to arch over the granite mass to the west, indicating that the mountain uplift is essentially a faulted arch.

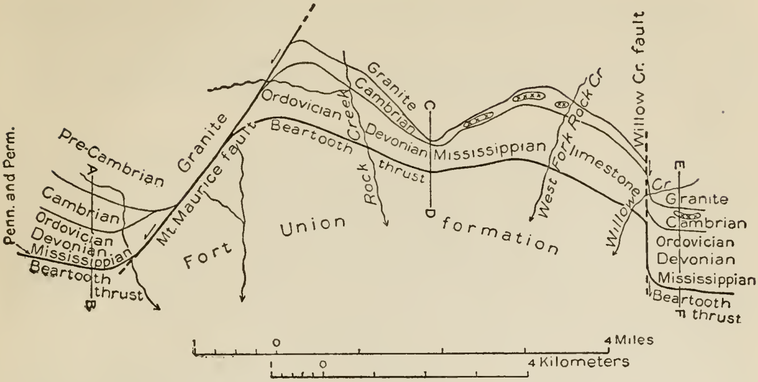

FIGURE 9.—Sketch map showing position of Mount Maurice and Willow Creek faults. (See also fig. 10)

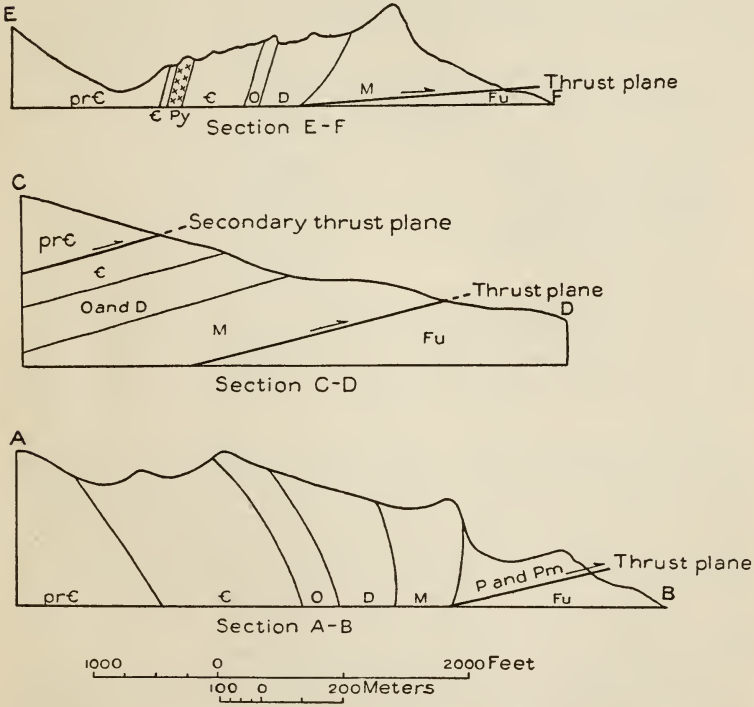

FIGURE 10.—Sections along lines A–B, C–D, and E–F, Figure 9. Py, Porphyry; prЄ, pre-Cambrian; Є, Cambrian; O, Ordovician; D, Devonian; M, Mississippian; P, Pennsylvanian; Pm, Permian; Fu .Fort Union

At the top of the switchback road are three natural bridges in the Madison limestone. This locality affords opportunity for fossil collecting and for study of the lithology of the Paleozoic formations and of the physiographic features of the Clark Fork Valley.

The route descends southeastward through the notch of Grove Creek to the road and thence back to Red Lodge, passing over the slightly flexed or undisturbed Wasatch and Fort Union beds at the foot of the vertical Paleozoic formations.

The beds exposed at the Eagle coal mine, 1½ miles (2.4 kilometers) up Foster Gulch from Bear Creek, consist of gently dipping Fort Union coal measures, and the locality is celebrated for fauna of primitive mammals.

These fossils are found in a thin layer of dark clay between the coal and the sandstone roof of the mine and apparently occur in the bed of an old stream or bayou, whose course is indicated by cone-shaped downward projections of the sandstone that may be fillings of spaces left by the roots of water plants. The fossils include, besides plant remains, the plates and bones of fishes, swamp turtles, *Champsosaurus*, and small mammals that are essentially similar to those found at Tiffany, Colorado, and at Big Sand Coulee in the vicinity of Polecat Bench, Wyoming. Collections of this fauna include the following mammals:

Planetetherium mirabile.	Labidolemur kayi.
Leptacodon siegfriedti.	Carpolestes nigridens.
Pentacodon inversus.	Carpolestes aquilae.
Protentomodon ursirivalis.	Thryptacodon pseudarctos.
Psittacotherium lobdelli.	Dissacus cf. D. navajovius.
Plesiadapis dubius.	Phenacodus primaevus.

The type specimens of these fossils are in part in the American Museum of Natural History, in New York City, and in part in the private collections of Dr. J. C. F. Siegfriedt, at Red Lodge, who was the first to bring this fauna to the attention of paleontologists.

Fossil leaves are found in considerable abundance in the beds underlying the No. 3 coal, and extensive collections have been made along Foster Gulch within a quarter of a mile (0.4 kilometer) of the entrance to the Eagle mine. In the gulch just below the mine portal fossil plants may be obtained from dark shales somewhat indurated by the heat of a dike, and this same ledge has also yielded insect and crustacean remains. More extensive plant collections have been obtained from beds about 100 feet (30 meters) below the No. 3 coal bed.

The Fort Union flora obtained in Foster Gulch is an assemblage of subtropical forms which grew in lowland swamps and presumably in a region of at least moderate rainfall. Laurels,

aralias, magnolias, figs, and poplars are the most abundant genera thus far identified in the collections.

RED LODGE TO BEARTOOTH BUTTE AND RETURN

By ARTHUR BEVAN and ERLING DORF

[Alternative with Red Lodge area excursion]

The 38-mile (61-kilometer) trip from Red Lodge, Montana, to Beartooth Butte, across the Wyoming line, takes the traveler from the foothills of the Beartooth Mountains to their summits. Geologically the traverse cuts deeply into the core of a huge asymmetric thrust-faulted anticline, exposing the complex mass of pre-Cambrian crystalline rocks, veneered by remnants of lower Paleozoic clastic rocks, limestones, and dolomites. The mountain upland reveals a great expanse of the subsummit peneplain, now standing 3,000 to 4,000 feet (914 to 1,219 meters) above the level of the plains, and upon this upland can be seen many evidences of an extensive ice cap, still represented by shrunken glaciers occupying cirques on the flanks of the Clark Fork-Yellowstone divide.

In Tertiary time erosion cut deeply into the pre-Cambrian crystalline complex in parts of the Beartooth Range and reduced large areas in the eastern part of the range to a surface of low relief (subsummit peneplain), which was seen, in profile, on the way to Red Lodge from the stopping points on Polecat Bench. Toward the canyon of Rock Creek, southwest of Red Lodge, the evenness of the plateau (peneplain) edge is again visible, especially toward the southwest.

Red Lodge is on the lowest of a series of well-developed Quaternary (Tertiary (?) in part) gravel benches, which slope gradually and evenly upward toward the mountain front. In traveling southwestward on this terrace the next higher bench may be seen to the northwest and still higher benches to the southwest. The westward-dipping Fort Union (Paleocene) sandstones, shales, and coal beds are visible along the valley slopes of Rock Creek, coated thinly with hetrogeneous boulders and gravel. Farther southwestward toward the mountains the valley of Rock Creek narrows considerably as the gradient increases toward the limestone "front," 4 miles (6.4 kilometers) from Red Lodge.

This easternmost face of the mountain mass is made up of nearly vertical to overturned Paleozoic limestones. The road crosses first the resistant Madison limestone of early Mississippian age. The less resistant Jefferson limestone (Devonian) forms the slight notch between the first ridge and the second,

which is made by the massive cream-colored Bighorn dolomite (Ordovician). The underlying Cambrian sediments (Gallatin limestone and Gros Ventre shales), because of their relative weakness, are not evident from the road, their position being indicated by grassy slopes between the Bighorn dolomite and the pre-Cambrian granite.

At road level the Paleozoic strata are slightly overturned to the northeast, so that their present dip is about 80° SW. North of the road between Rock Creek and West Fork, however, this overturning increases to 70° past the vertical.

Above the limestone scarp the creek has cut a deep valley into the core of pre-Cambrian crystalline rocks. Most of these rocks along the lower course of Rock Creek are pink granites and pegmatites, intruding gray granitoid rocks and gneisses.

The road crosses to the southeast side of Rock Creek 5½ miles (8.8 kilometers) from Red Lodge and enters the Beartooth National Forest just above the bridge.

The effects of glaciation in the main valley of Rock Creek are less marked than in some others. At its most vigorous stage (not the latest) the glacier built high lateral moraines, fragments of which remain almost down to the mouth of the canyon. The terminal moraine of this stage is eroded away. The floor of the mountain valley at this glacial stage was continuous with the Roberts Terrace, on the plains. This valley floor and terrace were trenched by Rock Creek about 200 feet (61 meters) deep before the less vigorous and extensive glaciation of the latest stage. Weak moraines of this last stage cross the present valley about 1 mile (1.6 kilometers) above the canyon mouth. Since the last glacial stage the stream has cut down less than 20 feet (6 meters). Above Richel Lodge, 12 miles (19 kilometers) from Red Lodge, is a compound lateral moraine which is the joint work of glaciers in the main valley and that of Lake Fork.

Although the valley of Rock Creek was doubtless much eroded by ice, the characteristic U form has been in large part lost, especially on the southeast side, by the easy sliding of the rocks, which has resulted in an almost uniform slope. Most of the valley floor is now covered by large alluvial fans. Torrential wash and landslides have so roughened the surface of these fans that in some places they simulate glacial moraines.

Because of the abundance and variety of glacial features in this area, a brief review of the glacial history is given here as an aid to the interpretation of the features to be seen. After uplift in late Tertiary and early Quaternary time an ice cap developed on the plateaulike summit of the Beartooth Range. This cap was mainly on the southwestward slope, but after the manner of ice caps it tended to spread in all directions. Some

AIRPLANE PHOTOGRAPH OF CIRQUES AND RESIDUAL GLACIERS ON CLARK FORK-YELLOWSTONE DIVIDE

Courtesy of Fifteenth Photo Section, Air Corps, U. S. Army. Remnants of peneplain in middle distance; stripped or "scrubbed" pre-Cambrian erosion surface in background. Looking southwest.

SOUTHEAST FACE OF BEARTOOTH BUTTE

Courtesy of Northern Pacific Railway. Photograph by Brown, St. Paul. Beartooth Lake in foreground; Cambrian shales and limestone in lower part of butte; massive Ordovician (Bighorn) dolomite at base of upper cliffs; thin-bedded Devonian limestones and shales in upper part of butte; lens of dark-red sediments in center of butte face.

AIRPLANE PHOTOGRAPH SHOWING PENEPLAIN REMNANTS ON CLARK FORK-YELLOWSTONE DIVIDE, FLANKED BY SURVIVING GLACIERS

Courtesy of Fifteenth Photo Section, Air Corps, U. S. Army. Stripped and glaciated pre-Cambrian erosion surface in the distance.

of the ice thus crossed the main northwest-southeast divide and descended in tongues toward the north and east. Thus the local supply of glacial ice in the basins of certain streams (Clark Fork, East Rosebud Creek, Stillwater River, and others) was reinforced by contributions from beyond these basins. At the time of maximum glaciation 15 extensive glacier systems existed in the northern and eastern parts of the range, some of them, especially those that were reinforced as stated above, passing out upon the plains beyond the mountain front.

Two stages of glaciation are commonly shown, and locally a third stage appears to be indicated. Many large cirques are present, especially in the higher parts of the range, and some of these still contain shrunken remnants of the once great glacial system. (See pl. 6.) A few of these residual glaciers are of unusual interest because they contain layers of ice in which myriads of grasshoppers are embedded. Examples of these are the two "grasshopper glaciers" at the heads of the West Fork of Rock Creek and West Rosebud Creek.

The trail to the summit follows the main fork of Rock Creek southwest from Richel Lodge and crosses an area of basic rocks which have been prospected for chromite. About halfway to the Rock Creek-Clark Fork divide the trail passes Mirror Lake, dammed by a morainic barrier. East of the lake is a large mass of porphyry, presumably of Tertiary age, which rises in precipitous slopes above the trail.

Close to the end of the switchback on the "slide-rock" trail, just before reaching the grassy uplands, there is a good exposure of the contact between the porphyry and pre-Cambrian amphibolite. A vertical pegmatite dike in the amphibolite is cut off sharply by the porphyry.

The steep ascent of the slide-rock trail finally brings the traveler upon the Line Creek Plateau, which is an extensive remnant of the subsummit erosion surface, characterized by gentle slopes and relatively low relief. Glaciation has affected this old surface differently on the two sides of the main divide. On the northeast side headward erosion has produced high, precipitous cirque walls, whereas the whole southwestern slope of the old surface has suffered scouring and grooving, with the development of numerous roches moutonnées, shallow rock basins, and glacial lakes. Several basic dikes traverse the country southwest of the divide.

A view southwest from the divide shows the broad old valley of Clark Fork, and beyond it cliffs of Paleozoic limestones and dolomites, overlain by volcanic rocks. This valley forms the dividing line between the Beartooth and Absaroka Mountains.

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The trail to Beartooth Lake descends to a valley trending slightly south of east, within which are exposed southwestward-dipping Cambrian quartzite, shale, and limestone downfaulted against the pre-Cambrian granite, which is exposed higher up in the south wall of the valley. The Cambrian shales and limestones crop out near the south valley wall and are somewhat fossiliferous, yielding mainly trilobites and calcareous algae. In the middle of the valley the basal Cambrian quartzite (Flathead) may be seen resting unconformably on the pre-Cambrian.

Beartooth Lake, at the base of Beartooth Butte (see pl. 7), is in a rock basin scooped by glaciation to the basal quartzite of the Cambrian and in places into the underlying granite. Northwestward across the lake the butte itself is seen to consist of two sets of formations, the lower dipping apparently 5° SW. and the upper lying essentially horizontal. Along the contact of the two series a trough filling of red sediments, which appears as a dark area in the center of the southeast face of the butte, is seen in cross section.

The lower grassy slopes rising above the lake consist of the Gros Ventre shales (Cambrian). The first line of cliffs at the base of the butte is made by the Gallatin limestone (Cambrian). The main cliff higher up commences with the thick Bighorn dolomite (Ordovician). Closer inspection shows the red beds above to lie, with a marked angular unconformity, in a rounded channel excavated in the Bighorn. The basal reddish beds consist of a conglomerate including fossiliferous limestone boulders, the whole giving the appearance of talus accumulations rather than an ordinary basal conglomerate. The upper beds in this trough are largely red shales and sandstones with a few thin beds of limestone, apparently of fresh-water origin. These reddish beds have yielded an ostracoderm fauna. The overlying limestone series which caps the butte is the Jefferson (Devonian).

The easiest ascent of the butte is made around the southwest side, following a longitudinal valley, and thence approaching the summit from the northwest and north. Considerable warping and faulting cause the beds to appear at different levels in the ridge to the northwest of the butte.

Physiographically the view from the summit northward shows (1) gently rounded grass-covered slopes ascending to the Rock Creek-Clark Fork divide after the fashion of greatly subdued mountains; (2) precipitous walls of cirques and glacial valleys indenting the old undulating summit surface; (3) several even-topped peaks and crest lines cut off by a plane (see pl. 8); and (4) uneven glaciated slopes leading southwest from the divide. To the northwest and west, across Clark Fork, the jagged peaks carved from the volcanic rocks of the Absaroka Range dominate

the view. Off to the southwest, however, are remnants of gently rolling grassy surfaces developed on the volcanic series, which appear to be comparable in nature and altitude to those seen on both sides of the backbone of the Beartooth Range.

This view shows an old erosion surface (subsummit peneplain), above which along the present main divide there rose at one time in the Tertiary a line of subdued mountain summits, some consisting entirely of pre-Cambrian granitic rocks and others possibly still capped by Paleozoic beds. After rejuvenation the whole area was deeply trenched and further modified by glaciation. Where the weak mantling sediments were last stripped from the peaks and divides the uncovered pre-Cambrian erosion surface was left with its exceptional flatness. These extensive flattish summit areas have also been interpreted as remnants of a summit peneplain.

Observations in the Beartooth Butte area extend the subsummit peneplain westward across the Clark Fork Valley and demonstrate its postvolcanic age, as remnants of it are to be seen on the volcanic rocks of the Absaroka Range.

RED LODGE TO BILLINGS BY WAY OF PRYOR GAP

By W. T. Thom, Jr., N. M. Fenneman, R. T. Chamberlin, and W. H. Bucher

[In part alternative with excursion from Pryor Gap to Sheridan]

The purpose of the excursion from Red Lodge to Billings is primarily to demonstrate the unique structure of the Pryor Mountains and secondarily to direct attention to the complex physiography of the region. The older Mesozoic beds are also well exposed along the route.

From Red Lodge the road to the Dry Creek oil field ascends to the Mesa Bench level and traverses its gently undulating surface, the original relief of which has been accentuated by recent downcutting of the old stream courses. This surface is the peneplain mentioned above as the highest of the terraces and correlated with Polecat Bench. Some of its faint depressions are gravel covered.

Six miles (9.6 kilometers) from Red Lodge there is a ridge overlooking the Dry Creek oil and gas field. From this ridge can be seen the dip of the strata flexed into the Dry Creek anticline, and farther east along the same structural line an area of Chugwater red beds exposed on the Red Top dome and also in the distance the median east-west monocline separating the southern part of the Pryor Mountains from the lower northern portion.

Westward the Dry Creek anticline, though largely obscured by terrace gravel, is essentially continuous with the Stillwater anticline, which is 30 miles (48 kilometers) in length, and converges westward with the Beartooth front, whose border of vertical Paleozoic limestones can be clearly seen from the vantage point above Dry Creek. Conspicuous slumps feature the slopes along the road as it descends the hill toward Dry Creek and point to the effectiveness of recent erosional activity.

At 14 miles (22.5 kilometers) from Red Lodge is the Dry Creek camp of the Ohio Oil Co., in the vicinity of which wells somewhat over 1 mile (1.6 kilometers) deep are producing oil from the Dakota(?) sands and gas from sands in the overlying Frontier formation. Slightly farther on, where the road passes through the southeast flank of the uplift, good exposures of typical lower Lance sandstones are to be seen just to the left of the road. It was from this general vicinity that fragments of dinosaur egg shells were obtained in 1930 by the Scott fund expedition under G. L. Jepsen.

Between this point and the irrigated lands along Clark Fork the southerly dip from the Dry Creek fold gives place to westerly dip from the Pryor uplift, and 24 miles (39 kilometers) from Red Lodge high sandstone ridges to the left (northwest) of the road are formed by the Parkman and underlying Eagle sandstones. Apparently the stream piracy by which the upper waters of Clark Fork were diverted into their present course has induced the recent downcutting so evident in the Dry Creek Valley.

A coal bed in the upper part of the Eagle has been mined here in a small way and has burned for a short distance along the outcrop, producing the streak of baked and reddened clays visible north of the mine.

A mile (1.6 kilometers) beyond the bridge over Clark Fork the road turns eastward from the Billings-Cody Highway, and 4 miles (6.4 kilometers) farther on it passes through a gap in the westward-dipping Frontier sandstone (Cretaceous). At 29 miles (47 kilometers) from Red Lodge a similar gap in the Mowry shale (Cretaceous) is passed. Eastward from points on the Mowry (Cretaceous) ridge there is a clear view of the Chugwater (Triassic) exposure marking the apex of the Red Top dome, flanked to the left by an escarpment of Jurassic beds and to the right by a ridge capped by the basal conglomerate of the Cloverly (Lower Cretaceous).

Still farther on the upper sandstone of the Cloverly makes a low ridge to the right of the road, behind which is a striking escarpment of Thermopolis shale capped by the resistant beds of the Mowry. A light-colored streak running along the midslope of this scarp marks the position of a sandstone (and associated

bentonite beds), which has been tentatively correlated with the upper sandstone of the Dakota of the Denver region.

On the Red Top dome, 34 miles (55 kilometers) from Red Lodge, is a well which, though drilled in search of oil, encountered artesian water in large volume. The water issuing from this well has in about 10 years eroded a channel 40 feet (12 meters) deep, 70 feet (21 meters) wide, and 1 mile (1.6 kilometers) long.

Excellent exposures of the upper Chugwater and lower Sundance are afforded by the escarpment north and east of the well, and complex faults are visible along the flanks of the uplift. Old oil seepages at the north side of the dome have caused local bleaching and saturation of beds in the upper part of the Chugwater, and near by are to be seen gypsum beds and a thin limestone, which lies just below the Triassic-Jurassic contact.

Eastward from the Red Top dome the rocks are broken into small blocks by a multitude of faults, whose effect is strikingly evident from a ridge of the basal conglomerate of the Cloverly formation, which is crossed by the road at a point 35 miles (56 kilometers) from Red Lodge. West and north of this ridge are fine scarp exposures of the Mesozoic formations, and in the more remote distance to the northwest and west the striking profiles of the Mesa Bench and the Beartooth subsummit peneplain can be seen.

About 2 miles (3.2 kilometers) southeast of the Cloverly ridge the Embar limestone makes the surface of the Black Butte dome. To the east can be seen the median (southern) east-west monocline of the Pryor Mountains, the entrance to the Pryor Gap, and southwest of the gap the old valley which probably was once traversed by the Shoshone River as it flowed across Polecat Bench and through the Pryor gorge (now a wind gap).

Beyond Bowler store, 38 miles (61 kilometers) from Red Lodge, the road affords a striking view of the abrupt faulted flexure by which the Madison limestone bends downward from the southern to the northern half of the Pryor Mountains.

The Pryor Mountains (see fig. 11) are peculiar in consisting of four blocks bounded on their northern and eastern margins by faults, partly exposed at the surface by sharp monoclinal flexures. Each block has been uplifted at its northeast corner. Figure 12 shows the position of the faulted monoclines which divide the mountains into four blocks, each sloping toward the southwest.

Along the mountain flanks small remnants of the Chugwater (Triassic) can be seen at the foot of the outer (Embar-Tensleep) dip slope, and within the inward-facing escarpment formed by the Embar and Tensleep is a second valley showing the red strata of the Amsden formation (Pennsylvanian).

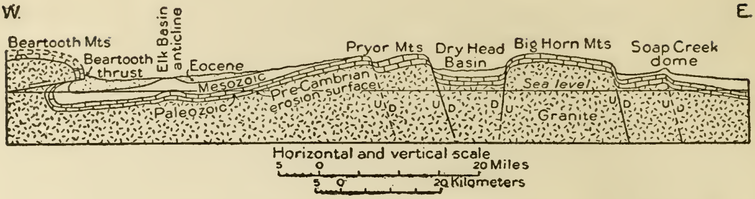

FIGURE 11.—Structure section across the Beartooth, Pryor, and Big Horn Mountains

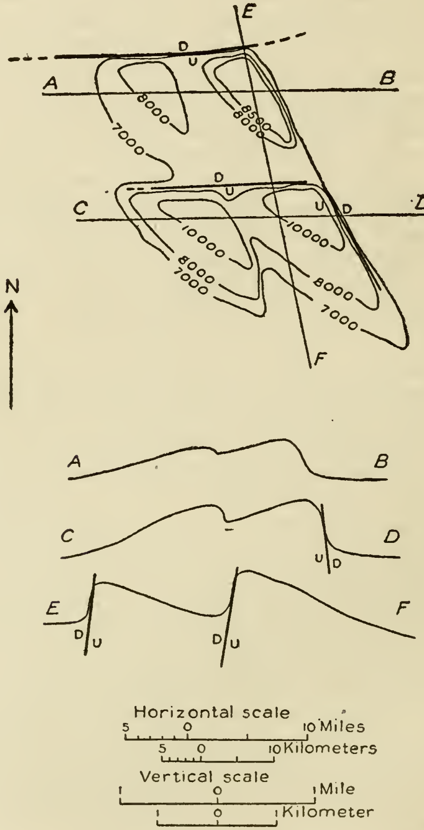

FIGURE 12.—Sketch map and sections showing structure of Pryor Mountains. D, U, Downthrown and upthrown sides of faults

The Pryor gorge (or wind gap) is carved through massive beds of Madison limestone and Bighorn dolomite (Ordovician). Just how deep the gorge once was in its lower reaches is not now ascertainable, as the lower slopes are now mantled by talus, and alluvial cones have been built in many of the tributary drainage lines.

At the outlet of the gorge, west of the road, strata are upturned by drag in the faulting of the northern monocline of the Pryor Mountains. Toward the east the formations visible are essentially horizontal, and the fault relationships are indicated by the relative position of the Carboniferous and Cretaceous beds—castles carved from the thick upper bed of the Madison limestone extending to the edge of the abrupt mountain front, whereas horizontal sandstone of upper Cloverly age lies at the foot of the scarp. Northward, along the left side of the creek, are rocky hills in which the Cloverly and Morrison are exposed, and beyond and northwest of Pryor post office the upper ledge of the Sundance (Jurassic), the overlying Morrison, and the basal conglomerate of the Cloverly can be seen arching over the apex of the Plum Creek dome.

At Pryor is a subagency of the Crow Indian Reservation, near which is the home of Chief Plenty Coos, whose biography, "American," by Frank Linderman, carries the reader back to the early 1800's, when, except for the Lewis and Clark expedition, the region was as yet untraversed by the white man, and the Absarokees (Crows) held it against hostile tribes.

The highest ridges visible north and northeast of Pryor are capped by the Mowry shale, and as the result of the prevailing gentle and undulating dip the Mowry and other sedimentary formations gradually descend northeastward and pass beneath the valley. Along the highway 1½ miles (2.4 kilometers) below Pryor is a gravel pit in which large limestone boulders derived from the near-by mountains are mingled with smaller cobbles and gravel composed of many kinds of igneous rocks characteristic of the Yellowstone-Absaroka volcanic area. The relative size of the smaller cobbles, their abundance within the Pryor Valley, and their absence from the divides to the west and east show that they were transported to their present position by a considerable stream (believed to have been the Shoshone River or Clark Fork, or the river resulting from the junction of the two) which once flowed from the volcanic area of the park eastward through the Pryor Gap.

Below Pryor there are on the right scattered exposures of the variegated clays and sandstones of the Cloverly formation, and farther on the Birdhead sandstone member of the Thermopolis shale forms a cap above dark Thermopolis shales in a series of

hills east of the road. Yet farther on grass-grown ridges of the Mowry shale converge to the creek, and 24 miles (39 kilometers) from Pryor cliffs of the Eagle sandstone rise from the valley. Near the site of the former railroad station of Coburn the Pryor Creek road joins the Hardin-Billings highway, which is followed to the top of the Yellowstone-Pryor Creek divide. Here is another extensive gravel-strewn plain which, in eastward succession, bevels without evident respect to hardness the Eagle sandstone, Claggett shale, Judith River formation, Bearpaw shale, and Lance formation. The gravel on the Yellowstone-Pryor Creek divide contains a considerable proportion of moss agate, quartzite, and other types of rock which are apparently absent in the gravel of the Pryor Creek Valley and are therefore evidently derived from a different source.

From the edge of this bench a series of bench or terrace levels, partly erosional and partly aggradational, can be seen southward in the Pryor Creek Valley, and here also there is a good view of the north end of the Big Horn Range, of the eastern Pryor monocline, and of Pryor Gap.

Good exposures of the Claggett shale (Upper Cretaceous) are afforded by the road excavations along the descent to the Yellowstone, and 98 miles (158 kilometers) from Red Lodge, at the Yellowstone bridge, cliffs of Eagle sandstone rise conspicuously on both the left and the right. The massive character of the Eagle in this locality throws light on the Upper Cretaceous paleogeography of the region, for here the Eagle, which is coal bearing in the Clark Fork Valley and areas farther west, passes into a strand facies which continues eastward for some 15 or 20 miles (24 to 32 kilometers), there merging into yellowish fossiliferous sandy shale of purely marine origin.

Because of the northward plunge of the Pryor Mountain uplift the Eagle sandstone has been eroded into a crescent of cliffs facing the irrigated valley on which Billings stands. From the vantage point on the rim north of Billings is afforded an excellent idea of the setting of the city, the significance of irrigation and agriculture being emphasized by the large beet-sugar refinery and meat-packing plant at the southern edge of the city.

Billings is at the center of radiating railway lines serving large regions that have extensive and diverse resources in soil, water, and minerals of many kinds, among which oil and gas, coal, clay, limestone, gypsum, and bentonite are of especial interest.

PRYOR GAP TO SHERIDAN

By R. T. Chamberlin, W. H. Bucher, and W. T. Thom, jr.

[Alternative with excursion Pryor Gap to Billings]

A section eastward across Yellowstone Park, the Big Horn Basin, Big Horn Mountains, Powder River Basin, and Black Hills uplift affords an opportunity to study typical details of Rocky Mountain structure and see how the character and intensity of the deformation varies from west to east.

Leaving Bowler, the road follows the old valley once occupied by the eastward-flowing Shoshone River and now by Sage Creek, which flows southward to join the Shoshone near its confluence with the Big Horn. Extensive gravel flats are present along the road, bordered on the east by terraced remnants of the old Polecat Bench plain and on the west by strike ridges of the Upper Cretaceous beds truncated at the Polecat level.

Near Warren, 13 miles (21 kilometers) from Bowler, remnants of the Polecat Bench are present on both sides of the road, and at the Montana-Wyoming line, 5 miles (8 kilometers) farther on, the derricks of the Frannie oil field are visible to the west of the road. Passing through Frannie, Deaver, and Cowley, the traveler crosses the Shoshone River just before reaching Lovell, 40 miles (64 kilometers) from Bowler. About 5 miles (8 kilometers) east of Lovell sharply flexed and faulted beds involved in the Sheep Mountain anticline can be seen south of the highway, and northward the two southern blocks of the Pryor Mountains (see fig. 12) tilted toward the southwest are well displayed.

Beyond the bridge over the Big Horn River just east of Kane, 52 miles (84 kilometers) from Bowler, the abruptness of the upturn at the mountain front is strikingly evident and is in marked contrast to the gentle westward dip of the Cretaceous beds along this part of the Big Horn Basin margin. An alternation of monoclines and flats is also observable in the mountain structure north of the road, and peneplaned surfaces on the mountains and terraces in the lower country are also evident. From vantage points along the winding road up the mountain, it is obvious that the Chugwater and Paleozoic formations are overturned and overthrust westward, the Chugwater red beds being brought into contact with the black Thermopolis shale and the Sundance, Morrison, and Cloverly formations having been cut out by the thrust. At the end of the steep ascent, as the top of the monocline is reached, the beds flatten abruptly, and for many miles to the east and southeast the road traverses

flat-lying Cambrian shales or the stripped surface of the pre-Cambrian, which combine to produce an area of low relief comparable to the subsummit peneplain on the Beartooth Mountains.

Farther northeast, in a direction at right angles to the strike of the range and of the monocline ascended by the road from Kane, the beds descend plainsward with moderate dips. The road, however, turns southeastward and leads into the greater central portion of the range. Here the border relations are reversed, and the west slope of the range is relatively gentle and the east slope is steep, with local overthrust toward the east. As this eastern flank of the mountains is approached, conspicuous cliffs of Bighorn dolomite rise above rounded slopes of the Cambrian shales, and where the road descends into the canyon there is a large landslide of Bighorn dolomite which has moved down the dip slope of the Cambrian. From this vicinity may be seen to the south a sharp local fold, overturned toward the west. As the road continues down the mountain side it winds back and forth along the strike of the strongly inclined Paleozoic beds and crosses the bright-red beds of the Chugwater at the foot of the slope. Bordering terraces similar to those in front of the Beartooth Mountains lie along the east flank of the Big Horn Mountains, and their gravel covering largely obscures the bedrock formations.

Within the valleys of the Tongue River and Goose Creek, near Sheridan, the coal-bearing strata of the Fort Union crop out. The striking red colors seen here and there along the highway have resulted from the spontaneous burning of the coal beds, which locally has produced great masses of natural brick along the coal outcrops. These valleys have long been the scenes of the extensive coal-mining operations which constitute one of the major industries of Sheridan. Farming by irrigation, stock raising, flour milling, and sugar refining operations are other important industries.

BIBLIOGRAPHY OF THE BEARTOOTH-BIG HORN BASIN REGION

1. Alden, W. C., Physiographic development of the northern Great Plains: Geol. Soc. America Bull., vol. 35, pp. 385–424, 1924.

2. Bartram, J. G., Elk Basin oil and gas field, Park County, Wyoming, and Carbon County, Montana: Structure of typical American oil fields, vol. 2, pp. 577–588, Am. Assoc. Petroleum Geologists, 1929.

3. Bevan, Arthur, Rocky Mountain front in Montana: Geol. Soc. America Bull., vol. 40, pp. 427–456, 1929.

4. Bevan, Arthur, Rocky Mountain peneplains northeast of Yellowstone Park: Jour. Geology, vol. 33, pp. 563–587, 1925.

5. Bevan, Arthur, Summary of the geology of the Beartooth Mountains: Jour. Geology, vol. 31, pp. 441–465, 1923.
6. Blackwelder, Eliot, Post-Cretaceous history of the mountains of central western Wyoming: Jour. Geology, vol. 23, pp. 97–117, 193–217, 307–340, 1915.
7. Chamberlin, R. T., A peculiar belt of oblique faulting: Jour. Geology, vol. 27, pp. 602–613, 1919.
8. Clapp, C. H., Bevan, Arthur, and Lambert, G. S., Geology and oil and gas prospects of central and eastern Montana: Montana Univ. Bull., Bur. Mines and Met. ser. 4, 1921.
9. Dake, C. L., The Heart Mountain overthrust and associated structures in Park County, Wyoming: Jour. Geology, vol. 25, pp. 45–55, 1918.
10. Dietz, C. S., The developed and undeveloped mineral resources of Wyoming: Wyoming Geol. Survey Bull. 21, 1929.
11. Fath, A. E., The age of the domes and anticlines of the Lost Soldier-Ferris district, Wyoming: Jour. Geology, vol. 30, pp. 303–310, 1922.
12. Hewett, D. F., The Heart Mountain overthrust, Wyoming: Jour. Geology, vol. 28, pp. 536–557, 1920.
13. Irwin, J. S., Faulting in the Rocky Mountain region: Am. Assoc. Petroleum Geologists Bull., vol. 10, pp. 105–129, 1926.
13a. Jepsen, G. L., Stratigraphy and paleontology of the Paleocene of northeastern Park County, Wyoming: Am. Philos. Soc. Proc., vol. 69, No. 7, pp. 463–528, 1930.
14. Robinson, E. G., Notes on Upper Cretaceous paleogeography: Am. Assoc. Petroleum Geologists Bull., vol. 8, pp. 554–559, 1924.
15. Simpson, G. G., Third contribution to the Fort Union fauna at Bear Creek, Montana: Am. Mus. Novitates No. 345, 1929.
16. Sinclair, W. J., and Granger, Walter, Eocene and Oligocene of the Wind River and Big Horn Basins: Am. Mus. Nat. History Bull., vol. 30, pp. 83–117, 1911.
17. Sinclair, W. J., and Granger, Walter, Notes on the Tertiary deposits of the Big Horn Basin: Am. Mus. Nat. History Bull., vol. 31, pp. 57–67, 1912.
18. Thom, W. T., Jr., and Dobbin, C. E., Stratigraphy of the Cretaceous-Eocene transition beds in eastern Montana and the Dakotas: Geol. Soc. America Bull., vol. 35, pp. 481–505, 1924.
19. Thom, W. T., Jr., The relation of deep-seated faults to the surface structural features of central Montana: Am. Assoc. Petroleum Geologists Bull., vol. 7, pp. 1–13, 1923.
20. Tomlinson, C. W., The middle Paleozoic stratigraphy of the central Rocky Mountain region: Jour. Geology, vol. 25, pp. 112–134, 244–257, 373–394, 1917.

APPENDIX A

PALEOCENE AND EOCENE FORMATIONS AND FAUNAS OF THE NORTHERN PART OF THE BIG HORN BASIN

By W. J. Sinclair and G. L. Jepsen

The Big Horn Basin is a broad syncline which lies within a horseshoe arc of high mountains—the Pryor Mountains on the northeast, the Big Horn Mountains on the east, the Bridger and Owl Creek Mountains on the south, and the Shoshone, Absaroka, and Beartooth Mountains on the west. Northward the basin merges with the plains of southern Montana. Sedimentary rocks of Cambrian to Upper Cretaceous age rest against the crystalline cores of these ranges and, descending with centripetal dip, pass beneath the sediments of the Cretaceous-Tertiary boundary, which occupy the center of the syncline.

The highest of these beds exposed to any notable extent is the *Systemodon-Coryphodon-Eohippus* zone of the Eocene Wasatch formation, which in some earlier reports has been called Gray Bull beds. (See table, p. 60.) It is best seen along the route of this excursion, where it forms the dissected lower and main slopes of the McCulloch Peaks, and again in the badlands that form the southwestern tip of Polecat Bench. It consists of reddish and bluish banded clays with lenticular sandstones of rapidly changing thickness intercalated at different levels. In contrast with the Eocene sediments seen in Yellowstone Park, this zone is predominantly nonvolcanic in character and represents fluviatile sands and muds and occasional palustrine deposits laid down in the sinking basin by aggrading rivers and streams.

The zone occupies much wider areas to the southeast of the route followed by this excursion. Where the Greybull and Big Horn Rivers and their tributaries cut across them, a multitude of creeks and nameless dry draws dissect its surface into magnificent and extensive badlands. There the largely bare surface of the feebly coherent rocks affords admirable opportunity for fossil collecting. This is a "bone digger's paradise." Fossil mammalian bones may be found almost anywhere, though not in great abundance at any one place. Even an experienced collector, however, may walk for several days without finding anything of importance. As a result of years of intensive collecting, the clays (good specimens are rarely

found in the sandstones) have afforded a large fauna of lower Eocene mammals, two giant flightless birds, crocodiles, fresh-water fishes, and both fresh-water and terrestrial mollusks. This is the "*Coryphodon* zone" of Osborn, regarded as correlative with the Sparnacian Stage of Europe and characterized by the first appearance in North America of perissodactyls, paramid rodents, and adapid primates. Some of the characteristic genera are as follows:

Eohippus, the "dawn horses," with many species.
Homogalax, the rhinoceros-tapir ancestor, several species.
Phenacodus, a large condylarth, several intergrading species.
Hyopsodus, a small condylarth, formerly described as an insectivore.
Sinopa, *Pachyaena*, *Oxyaena*, *Palaeonictis*, creodonts or primitive flesh eaters. Several additional and less abundant genera also occur.
Pelycodus and *Tetonius*, primates referable to the Adapidae and Tarsiidae, respectively.
Coryphodon, an amblypod, the largest mammal from the lower Eocene.
Paramys, a large rodent.
Esthonyx, a tillodont.

The fossil collection par excellence, with most of the types from this formation, is in the American Museum of Natural History in New York City. A smaller collection, with a number of interesting and unique types, is on exhibition in the Geological Museum of Princeton University, Princeton, New Jersey. A still smaller collection is at Amherst College, Amherst, Massachusetts.

The collections contain less than a dozen mounted skeletons in all. Most of the material consists of single teeth, jaw fragments with a few teeth, or maxillae and lower jaws more or less complete and in association. The fossils are mostly heavily invested with concretionary material, either calcareous, ferruginous, or both. Many of the bones show evidence of considerable decay before fossilization and are usually distorted by rock pressure. They are a dark chocolate-brown to black in color, unless stained bluish by vivianite or showing a whitish patina from weathering.

The base of the Eocene sediments is seen in the southeast wall of Polecat Bench. It is the line that separates the red-banded beds of this series to the southwest from the drab Paleocene beds which occupy the center of the escarpment. Closer inspection proves the contact here to be conformable and transitional, though farther out toward the edge of the basin the same contact is marked by angular unconformity, with Eocene beds truncating the uplifted Paleocene strata. This indicates that after the basin structure had become accentuated, at the end of Paleocene time, sedimentation continued within the basin while the upturned margins suffered erosion.

Like the Eocene, the Paleocene sediments, known as the Fort Union formation, are all of fluviatile and palustrine origin. They comprise in this region at least four mammalian faunas. The upper two of these faunal horizons are best represented in

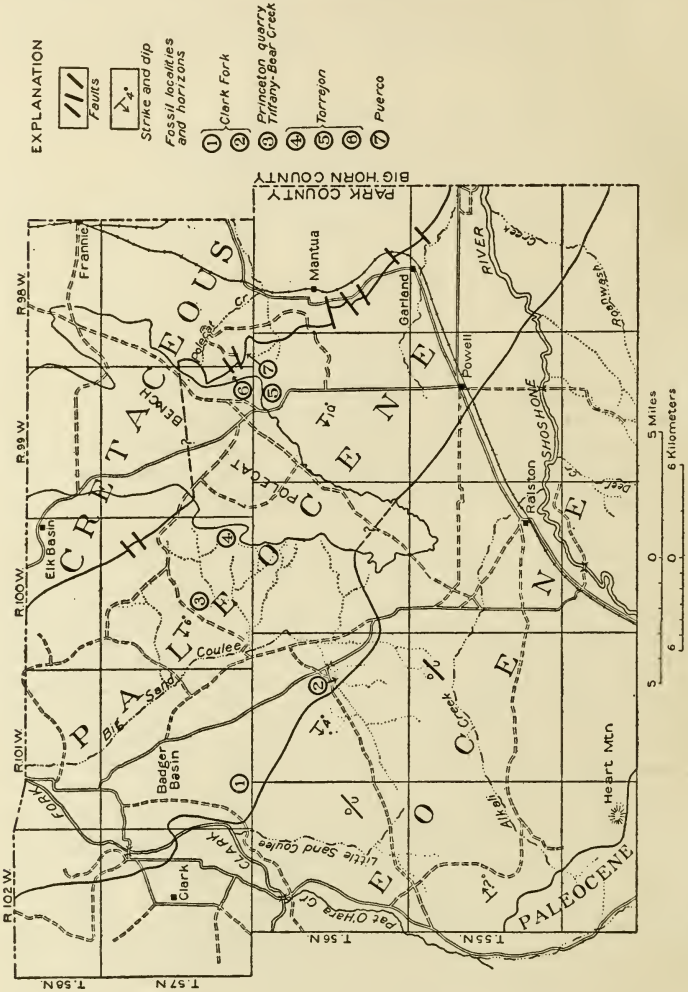

FIGURE 13.—Geologic sketch map showing vertebrate fossil localities in Polecat Bench area. (By G. L. Jepsen)

the vast region of badlands to the west of Polecat Bench (not visible from the main motor road) carved by Big Sand Coulee and the maze of tributaries that lead northwestward into Clark Fork. (See localities 1 and 2, fig. 13.) Here, immediately

underlying the banded Eocene clays of the Gray Bull beds with *Eohippus*, lie the Clark Fork beds, for which this is the type locality. Many of the genera and even certain of the species of the Clark Fork beds extend upward into the Eocene, but no perissodactyls have been found in them.

Lower down in the Fort Union formation (locality 3, fig. 13) occurs a fauna which is the equivalent of the Tiffany beds of Colorado, described by Granger in 1917. It comprises four multituberculates, three insectivores, two plesiadapids, two condylarths (*Phenacodus* and *Ectocion*), an artiodactyl, and the highly problematic form *Carpolestes*. This is also the fauna which occurs in the roof shales of the Eagle mine at Bear Creek, Montana. (See itinerary, p. 38.)

Where the motor road ascends the steep southeastern face of Polecat Bench (locality 5, fig. 13), a still lower faunal horizon is represented by teeth or jaw fragments of *Titanoides*, *Claenodon*, *Tetraclaenodon*, and *Plesiadapis*, besides bones of the rhynchocephalian *Champsosaurus*. This fauna is definitely correlative with the Torrejon of New Mexico and establishes fully the Paleocene age of this lower part of the Fort Union formation. Besides these sparse remains of mammals and reptiles, some limy layers contain shells of *Unio*, *Vivipara*, *Lioplax*, and *Goniobasis*, and occasionally erect silicified tree trunks are found.

The lowest part of the Fort Union formation and its contact with the underlying Lance formation are well exhibited in the amphitheaterlike embayment in the east face of Polecat Bench formed by the headwaters of Polecat Creek. A sandstone in the upper part of the exposures at this point (locality 6, fig. 13), which are described in greater detail on page 35, has furnished a mammalian fauna of Torrejon age with *Claenodon*, *Tetraclaenodon*, *Chriacus*, *Ptilodus*, *Palaeosinopa*, *Tricentes*, and *Plesiolestes problematicus* Jepsen, of which the first four are the commonest elements. Stratigraphically this fauna lies below that described above.

At the base of the somber-colored shales and sandstones with coal seams lies a massive cliff-forming yellowish-brown sandstone almost 130 feet (40 meters) thick. In the basal layer of this sandstone a mammal fauna was found comprising such forms as *Eucosmodon gratus* Jepsen, *Loxolophus nordicus* Jepsen, *Eoconodon* sp., *Oxyacodon* sp., and *Conacodon* sp. This fauna is correlative with that of the Puerco beds of New Mexico.

At 20 feet (6 meters) below the base of this sandstone dinosaur bones have been found. The horizon is that of the Lance formation. The base of the sandstone, then, marks the lower contact of the American Paleocene, at the much debated transition between Cretaceous and Tertiary. This is the boundary between Mesozoic and Cenozoic as recognized by American vertebrate paleontologists.

Paleocene and basal Eocene deposits of northern Big Horn Basin

[In part after Simpson]

Age		Some characteristic faunal elements	European time equivalents
Lower Eocene.	Graybull.	First perissodactyls, rodents, and adapids; last Champsosaurus.	Sparnacian.
Marked faunal change—			
Paleocene.	Clark Fork.	Many lower Eocene genera, but no perissodactyls or rodents. Champsosaurus, Carpolestes, etc.	Thanetian. Conglomerate of Cernay.
	Tiffany-Bear Creek.	Mostly a microfauna. Plesiadapis, Carpolestes, Phenacodus, etc. First artiodactyls.	
	Torrejon.	Ptilodus, Chriacus, Claenodon, Tetraclaenodon, Titanoides, Champsosaurus.	None known.
	Puerco.	First carnivores and ungulates. Taeniolabis, Triisodon, Ectoconus, Conacodon, Champsosaurus.	
Great faunal change—			
Upper Cretaceous.	Lance.	The last of the dinosaurs. Champsosaurus. Mammals: Insectivores, didelphids, multituberculates.	Danian.

APPENDIX B

HISTORY OF PETROLEUM DEVELOPMENT IN THE BIG HORN BASIN REGION

By Wilson B. Emery

Next to the development of water supplies for irrigation the exploitation of petroleum and natural gas in the Big Horn Basin and its northward continuation has, perhaps, exerted the most important single influence upon the settlement and commercial development of this region.

Oil seeps at several points in the Big Horn Basin were known to the pioneers, but active prospecting for oil was not undertaken until the first decade of the twentieth century. The earliest commercial discovery was made in 1906 in the Garland field, near the town of Byron. In the years immediately following several shallow wells yielding small quantities of high-gravity crude oil were completed, the product being refined in a small plant located at Cowley. Prospecting was also under way elsewhere in the basin during this period, and some additional discoveries were made, notably at Greybull, but no richly productive area of large size was revealed until 1914, when the productive Grass Creek field, 40 miles (64 kilometers) southeast of Cody, was brought in. The light oil from this new field found a ready market and added impetus to the search for oil in other parts of the Big Horn Basin and the adjoining areas in Montana, which resulted in the discovery in 1915 of the Elk Basin oil and gas field, on the Wyoming-Montana line. (See pl. 5.) The discovery was quickly followed by drilling in the Montana part of the field, which resulted in the first commercial production of oil from that State. All this light oil produced was obtained from sands near the base of the Upper Cretaceous.

Meanwhile prospecting for the black oils found in the Carboniferous of the region was also in progress, and in 1917 the Warm Springs field, east of Thermopolis, was discovered. Other black-oil fields were also discovered, at about this time and subsequently, along the borders of the Big Horn Basin, but to date no black oil has been produced in the Montana portion of the area.

With improvement in drilling equipment deeper drilling in already productive gas and light-oil fields was undertaken in

Wyoming, and several of these fields have been found to be underlain at depth by extensive black-oil pools. The first of these pools discovered was that at Grass Creek, in 1920. The large Oregon Basin field was developed in 1927, and black oil was discovered in the deep sands of the Byron and Garland fields in 1930.

In Montana wildcatting showed increased activity after the discovery of the Elk Basin field, but only two oil fields have been discovered in the area covered by this summary. Of these the Lake Basin field was found in 1924, and the Dry Creek field in 1930.

A total of 38,613,333 barrels (6,130,252,747 liters) of oil had been marketed from the Big Horn Basin, Dry Creek, and Lake Basin area to January 1, 1931, of which about 83 per cent was light oil, and of this total 233,472 barrels (33,286,015 liters) had been marketed from the Dry Creek and Lake Basin fields of Montana.

The light oil produced in this general area is derived from the Frontier and Dakota (?) sands, of Upper Cretaceous age; black oil is found in the Embar (Permian), Tensleep and Amsden (Pennsylvanian) formations, and the Madison limestone (Mississippian). Minor amounts of black oil have been discovered in the Chugwater red beds of Triassic age, but the Morrison (Lower Cretaceous (?)) and Sundance (Jurassic) formations have not yielded oil in this area.

There is a possibility that future drilling may uncover additional light-oil pools in the Big Horn Basin, Dry Creek, and Lake Basin region, and if experience is a guide additional black-oil pools will doubtless be discovered. However, it will probably be some years before extensive prospecting in this area is again undertaken, especially for black oil, of which a very large shut-in reserve is already developed awaiting a market.

APPENDIX C

GRAVITY DATA

By William Bowie

Figure 14 shows the location of the stations so far occupied by United States and Canadian geodetic surveys in this general region and the anomalies obtained as computed by the Hayford method.

Two features stand out as of distinct interest. The area of the Black Hills is indicated by a group of positive anomalies inclosed by the boot-shaped contours marked "+40." Extending from it in a northwesterly direction are two belts of positive anomalies inclosed by the "+20" contours. It may be significant that thinning of stratigraphic units in this general region and what is known of the structural trend in the pre-Cambrian basement (see pl. 3) suggest that here the basement rocks lie nearer the surface than on each side. This may indicate a relation between the relatively high positive anomalies and the nearness to the surface of the pre-Cambrian rocks due to broad uplifts.

The Rocky Mountain front in Montana and beyond, on the other hand, so far as it appears on this map, shows low anomalies, all with one exception deviating less than 0.020 degree from zero in either direction. This is in harmony with observations in other intensely disturbed regions with sharp folding and thrustfaulting.

The three stations in Yellowstone Park show the highest and the lowest anomalies of this portion of the Rocky Mountain front lying side by side, a circumstance which may possibly bear a relation to the very recent magmatic movements in this region.

Additional stations are now being established east of Yellowstone Park in order to test further the relation of geodetic to geologic observations.

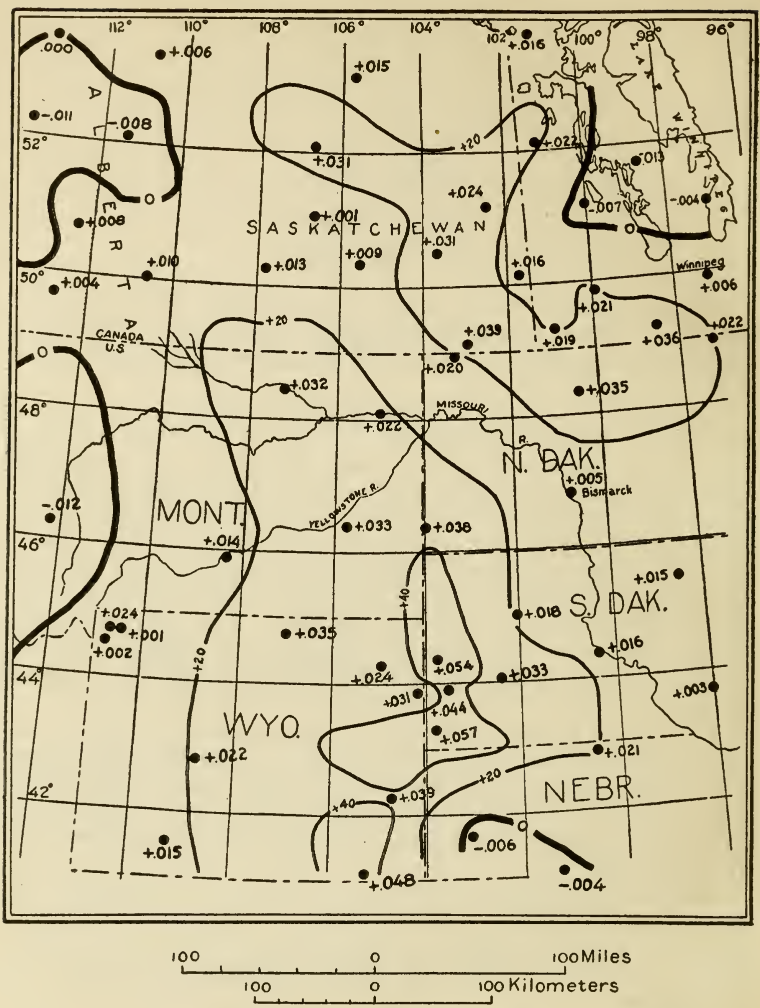

FIGURE 14.—Gravity anomalies in the Yellowstone-Beartooth-Big Horn and adjacent regions. The anomalies result from an isostatic reduction in which a depth of 70.6 miles (113.7 kilometers) is used for the thickness of the earth's crust. An anomaly of 0.001 is about one millionth of the average value of gravity and may be expressed in dyne or cm/sec² of acceleration